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Contents

Vol. 9, No. 2, 2001

Simultaneous English language translation of the journal is available from MAIK "Nauka/Interperiodica" (Russia).
Stratigraphy and Geological Correlation ISSN 0869-5938.

C-Isotope Chemostratigraphic Correlation of the Upper Riphean in the Urals and Arctic Region
M. E. Raaben 107

On Possibility of Using the Global Permian Stage Scale in the Tethyan Region
E. Ja. Leven 118

Biostratigraphy of the Marine Paleogene in the West Siberian Plate
M. A. Akhmet'ev, G. N. Aleksandrova, E. O. Amon, V. N. Beniamovskii, E. M. Bugrova, O. N. Vasil'eva, Z. I. Glezer, V. I. Zhelezko†, N. I. Zaporozhets, G. E. Kozlova, I. A. Nikolaeva, T. V. Oreshkina, L. A. Panova, E. P. Radionova, N. I. Strel'nikova, and A. I. Yakovleva 132

The Paleogene Reference Section of the Kril'on Peninsula (the Kitosiya River, Southern Sakhalin)
L. I. Fot'yanova, M. Ya. Serova†, V. G. Gal'versen, A. E. Zharov, N. M. Grokhotova, and V.P. Tuzov 159

The Cenozoic of Sakhalin: Modern Stratigraphic Schemes and Correlation of Geological Events
Yu. B. Gladenkov 177

Stratigraphic and Geographic Distribution of Late Eocene Foraminifers in the Northern Periphery of the Tethys
E. M. Bugrova 191

Transgressive and Regressive Seismic Facies of the Danube Deep-sea Paleofan
V. N. Moskalenko 204

Stratigraphy and Geological Correlation, Vol. 9, No. 2, 2001, pp. 107–117. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 9, No. 2, 2001, pp. 3–14.
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C-Isotope Chemostratigraphic Correlation of the Upper Riphean in the Urals and Arctic Region

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Abstract—Two secular variation curves of C-isotope composition in the Late Riphean carbonates are discussed. The curve approximating data on the Upper Riphean deposits of Spitsbergen, Greenland, northwestern Canada, and southwestern Africa (Kaufman and Knoll, 1995) is now accepted as the model one for the second half of this stratigraphic interval and as useful for global correlations. The careful analysis of the used chronometric calibration of that curve shows that it is vulnerable in the lower, relatively older part of the curve. Comparison of the model curve with another one recently obtained for the Karatau Group, the Upper Riphean stratotype in southern Urals (Podkovyrov *et al.*, 1998) demonstrate that they are of the in-phase character. They both depict the identical number of large positive and negative peaks despite the lower amplitudes of $\delta^{13}C$ variations in the Uralian section that is conditioned, in general, by the lower values of extreme positive points. The in-phase character of two variation curves suggests, along with biostratigraphic, isotopic-geochronological and regional geological data, that the model one well characterizes not only the second half, but also the older part of the Late Riphean time. The available difference in configuration of uppermost parts of the curves under comparison appears to be a consequence of the Upper Riphean stratotype incompleteness, as there is a large pre-Vendian stratigraphic hiatus and another one below the Uk Formation, which was underestimated in previous works.

Key words: Neoproterozoic, Upper Riphean, Karatau Group, chemostratigraphy, C-isotope correlation, southern Urals, Spitsbergen.

INTRODUCTION

Perspectives of using secular C-isotope variations in carbonates for the Upper Precambrian stratigraphy are discussed in many recent publications. In addition to consideration of such variations in the Proterozoic–Cambrian boundary beds (Brasier *et al.*, 1990, 1993; Fairchild *et al.*, 1990; Kirschvink *et al.*, 1991; Magaritz *et al.*, 1991; Kaufman *et al.*, 1996), a serious attention was paid to the Late Riphean and Vendian intervals. The synthesis of data on a series of sections in the western Arctic regions (Knoll *et al.*, 1986; Fairchild and Spiro, 1987; Kaufman *et al.*, 1992; Narbonne *et al.*, 1994), equatorial Africa (Kaufman *et al.*, 1991), Asia Minor (Burns and Matter, 1993), and southern China (Lambert *et al.*, 1987) was recently used to plot the generalized curve of secular $\delta^{13}C$ variations in the Neoproterozoic (Kaufman and Knoll, 1995). This curve attributed to two younger Late Neoproterozoic subdivisions, which correspond to the second half of the Late Riphean and Vendian of the traditional Russian scale, is now accepted as the model one for this part of the geological record. The data on Riphean sections in various regions of northern Eurasia appeared almost at the same time (Pokrovskii and Vinogradov, 1991; Semikhatov *et al.*, 1995; Knoll *et al.*, 1995; Podkovyrov and Vinogradov, 1996; Podkovyrov *et al.*, 1997; Podkovyrov, 1997) and to a considerable extent supplemented C-isotope characteristics of the terminal Middle and early Late Riphean. Soon afterward, Podkovyrov *et al.* (1998) published the complete investigation results on the Upper Riphean stratotype, the Karatau Group of southern Urals. As it is rightfully stressed in the last work, these results elucidate the Upper Riphean chemostratigraphy established in one type succession. It is also important that the studied samples were selected according to modern criteria of selecting the least altered rocks, and the analytical results obtained are undoubtedly valuable from this standpoint. Data on the Uralian succession, when compared with those used in the construction of the model curve (Kaufman and Knoll, 1995), demonstrated that the $\delta^{13}C$ variations in the Karatau Group are of a lower amplitude, but fit in general the model variations of this parameter. It was also noted for the Urals that the lengthy positive excursions of $\delta^{13}C$ values are missing from the interval of the upper Upper Riphean, and this was assumed to be an indication that the Uralian succession of carbonate formations is stratigraphically incomplete because of cryptic hiatuses, which have not been recognized in the Karatau Group (Podkovyrov *et al.*, 1998). With reference to the Upper Riphean type section, which is considered as lacking breaks by many researchers, the noted problem is indeed important.

There is as well another more general question: to what extent reliable is the time calibration of the model

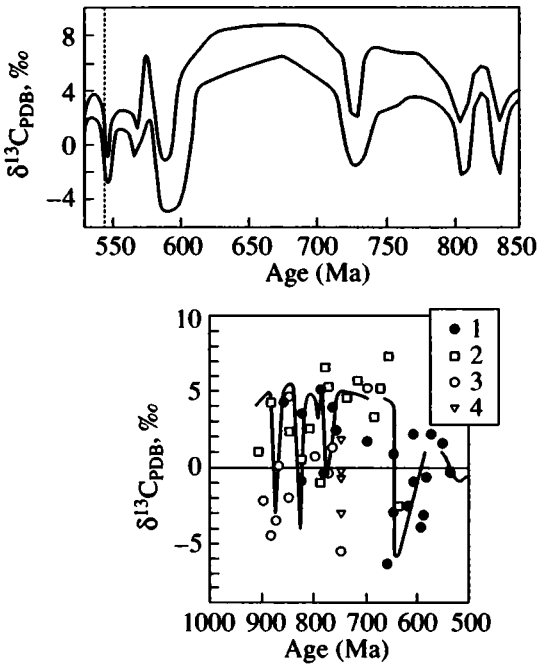

Fig. 1. Model curve (upper diagram) of secular $\delta^{13}C_{carb}$ variations during the Neoproterozoic (Kaufman and Knoll, 1995) and original plot (lower diagram) of the same variations in the Arctic Upper Riphean deposits (Spitsbergen and Greenland; data after Knoll *et al.*, 1986): (1) average for five and more analyses; (2) average for two–four analyses; (3) one analysis; (4) data on rocks of presumably nonmarine genesis.

$\delta^{13}C$ variation curve during the Late Riphean? This question is unavoidable owing to discordance of two variation curves, when they are plotted in the identical manner. Actually, the curves under discussion are published on different scales: in the model curve, $\delta^{13}C$ values are plotted versus the time scale (Fig. 1) and variations of this parameter in the Uralian section are considered versus the thickness scale. In addition, the picturing modes of minor variations are mutually inconsistent, and all this masks the real relationships between the two curves. A somewhat better correspondence can be seen, when the Uralian curve (Fig. 2A) is compared with the curve characterizing carbonates from Spitsbergen and Greenland (Fig. 2B), because they are plotted in a similar manner.

In Fig. 3, original data of the model and Uralian curves are plotted in the identical manner. Since minor fluctuations of the model curve are bounded by two contours corresponding to their minima and maxima, the Uralian curve is simplified in the same style, so that we can compare only the long-term $\delta^{13}C$ variations. The interregional correlation of short-term variations during the Neoproterozoic, like during the Cambrian, always meets serious difficulties (e.g., Kaufman *et al.*, 1997; Hoffman *et al.*, 1998).

MODEL AND URALIAN $\delta^{13}C$ VARIATION CURVES: SIMILAR AND DISTINCTIVE FEATURES

The model curve, as is mentioned above, represents the compilation of data from several regions. Principal data characterize succession of the Spitsbergen Archipelago, and others only supplement them (Kaufman and Knoll, 1995). Earlier, Knoll *et al.* (1986) published the $\delta^{13}C$ variation curve based on data for the western Spitsbergen, Nordaustlanded Island, and eastern Greenland. This curve resembles the present-day model curve in configuration despite the different calibration versus the time scale (Fig. 1). Below, I consider only the pre-Vendian part of the model curve, because the Uralian data do not characterize the Vendian lacking carbonate rocks in the southern Urals. This part of the model curve, according to its authors, is limited by the Cryogenian (initial boundary 850 Ma old), the two subdivisions of which are separated by the chronometric boundary of 730 Ma. The Cryogenian terminal boundary (in other words, the Vendian base) is placed at the occurrence level of the Varanger tillites and corresponds to the time mark of 590 Ma chosen on the basis of dates from Massachusetts and Newfoundland, where these tillites overlie volcanic rocks with U–Pb zircon ages of 602 ± 3 and 606 ± 2.4 Ma (Kaufman and Knoll, 1995, and references therein). In Europe, this level is accepted as corresponding to the older date of 650 ± 20 Ma (see details in Semikhatov *et al.*, 1991), which seems to be overestimated in the light of recent results.

The Uralian curve depicts $\delta^{13}C$ variations in the greater portion of the Upper Riphean type section represented by carbonate succession of the Karatau Group overlying basal siliciclastic deposits of the Zil'merdak Formation. Carbonates may represent up to 90% of this succession, but its middle interval corresponds to the siliciclastic Inzer Formation widespread in the western slope of southern Urals. Only in the western periphery of the Bashkirian meganticlinorium, carbonates are dominant rock types in the succession above the Zil'merdak level, being interrupted in the middle by three minor members of siliciclastic rocks. The base of the lower and the top of the upper members correspond respectively to the base and top of the Inzer Formation distinguished in eastern sections. In this connection, some researchers (Kozlov *et al.*, 1991) consider the term "Inzer Formation" as outlining the middle interval of the Karatau Group even in those areas, where the group predominantly consists of carbonate rocks (Fig. 2A). It seems to be more correct to term this interval as the regional Inzer Horizon or Superhorizon (Komar, 1978; Raaben and Komar, 1983). The mentioned siliciclastic members of variable thickness are intermittent in lateral directions, and the Karatau Group sections always reveal some peculiar features in their western distribution area (Fig. 4).

The composite stratigraphic section, used as the basis for studying $\delta^{13}C$ variations in the Karatau Group,

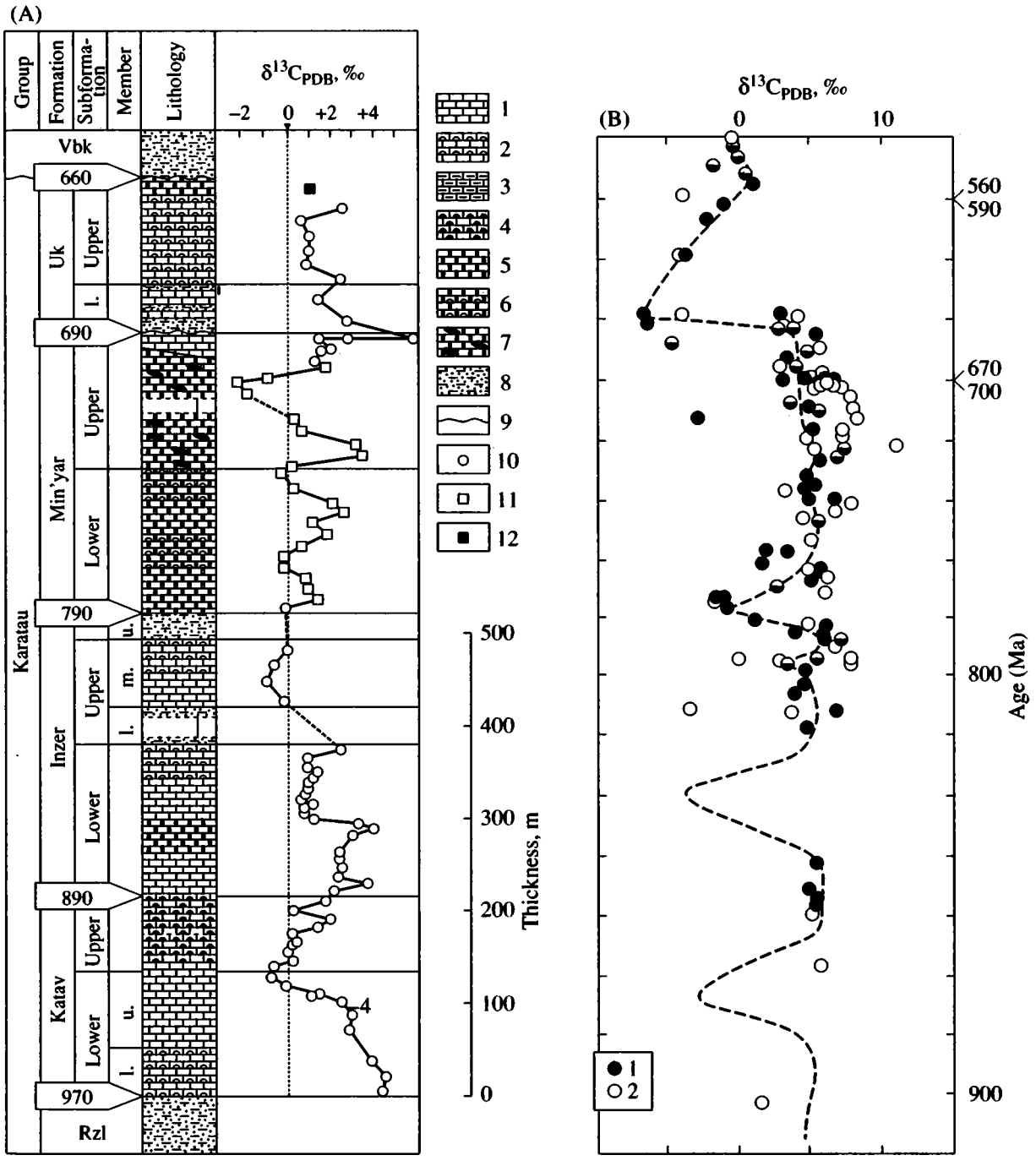

Fig. 2. $\delta^{13}C_{carb}$ variations (A) in the Upper Riphean Karatau Group of southern Urals and (B) in the Middle Hecla Hoek Supergroup of Spitsbergen (Podkovyrov *et al.*, 1998; Knoll *et al.*, 1986). Symbols for the diagram (A): (1–4) micrite-clastic (1), stromatolitic (2), clayey (3), and striate-laminated limestones; (5–7) stromatolite-free (5), stromatolitic (6), and cherty (7) dolomites; (8) siliciclastic rocks; (9) unconformity; (10) data on unaltered limestones; (11) data on unaltered dolomites; (12) data on altered dolomites; (Vbk) Bakeevo Formation of the Vendian; (Rzl) Zil'merdak Formation of the Upper Riphean (inset figures denote the estimated isotopic age, Ma). Symbols for the diagram (B): (1) more than four coupled C_{carb} and C_{org} analyses; (2) four or less coupled analyses.

corresponds in its main part to the section of the Adzhigardak anticline in the western zone of the Bashkirian meganticlinorium. The most remarkable feature of this well-studied section exposed near the town of Min'yar is the minimum thickness of lower and upper siliciclastic members in contrast to the middle one that is rather thick (110 m). In the presented lithostratigraphic column (Fig. 2A), this member is designated as the lower one in the upper Inzer Subformation (in accord with data of Kozlov *et al.*, 1991), although it was earlier

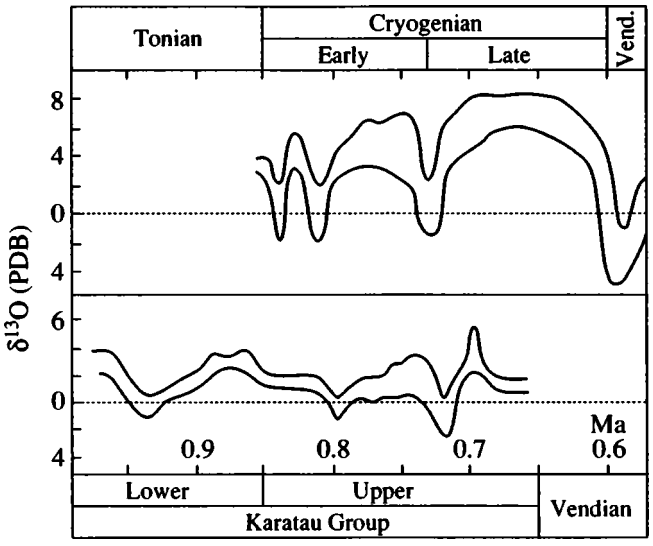

Fig. 3. Unified curves of $\delta^{13}C_{carb}$ variations during the Late Riphean: Cryogenian segment of the model Neoproterozoic curve (above) and simplified Uralian curve (below).

described and mapped as the whole Inzer Formation (Bekker, 1961; Raaben, 1975; Raaben and Komar, 1983; Raaben and Kozlov, 1983; Keller and Raaben, 1989). C-isotope characteristics for the Uk Formation, uppermost in the Karatau Group, and for the terminal beds of the Min'yar Formation were obtained on the basis of samples from the western limb of the Suleimanovo anticline near the Shubino Settlement, where the respective sections are more representative than near the town of Min'yar. The minimum age limit of the Karatau Group is 660 Ma, and the base of its carbonate rocks is considered to be ca. 970 Ma old (Fig. 2A).

In the comparative plot (Fig. 3), the model and Uralian curves reveal, despite the different time-range, the identical number of large $\delta^{13}C$ anomalies: four positive maxima divided by three negative minima. Also remarkable is similarity of both curves in their middle parts, where positions of the second and third negative excursions are almost coincident and configurations of separating positive anomalies are comparable despite their difference in absolute values. According to Podkovyrov *et al.* (1998), difference between the curves is most distinct in their terminal parts (Fig. 3), where the model one shows the lengthy maximum of positive values, which extends to the Vendian base in contrast to the narrow peak ca. 690 Ma old in the Uralian curve.

As for the curve segments older than 800 Ma, two positive peaks noticeable in the segments and depression between them are asynchronous and different in shape. In the Uralian curve, these elements characterize the lower Karatau Subgroup, i.e., the time interval of more that 100 Ma, whereas in the model one, they are located within the initial Cryogenian and appear to be about 50 Ma long. Taking the model curve for the global standard at its face value, one should admit that three large excursions of $\delta^{13}C$ values are missing from the Early Cryogenian part of the Uralian curve. To properly explain this discrepancy, we should evaluate first the validity of the distinguished successive excursions and, second, the correctness of their positions on the chronometric scale accepted in the model curve.

The excursions under consideration appear to be well substantiated. Actually, the data used to plot the $\delta^{13}C$ secular variations are obtained in several separate regions, but the succession basic in this case is that of the Upper Riphean rocks exposed in the Ny-Friesland Peninsula of western Spitsbergen, where it is uniquely thick and complete. This succession corresponds to the Middle Hecla Hoek Supergroup, the upper subdivision of which (Academikerbreen Group) is composed almost entirely of carbonate sediments. In the lower subdivision (the Veteranen Group), carbonates are less abundant than siliciclastic deposits, and the respective segment of the $\delta^{13}C_{carb}$ variation curve is depicted on the basis of analogy with the $\delta^{13}C_{org}$ values measured in shales. Data on the adjacent Nordaustlanded Island prove the validity of the established curve, as the composite one replicates it against the background of a wider diapason of minor fluctuations (see Fig. 1). Data on other, more or less fragmentary and separated sections were used to substantiate in principle the global significance of the obtained model curve. Some of these sections also yield information important for the geochronometric calibration of the established secular variations of $\delta^{13}C$ values.

Such a calibration is indeed of vital significance. As one can see in the graph (Fig. 3), the curve takes origin at the beginning of the Cryogenian (850 Ma), and two closest negative excursions are accepted to be 800 and 730 Ma old. None of these dates is substantiated in the isotopic-geochronological aspect, because the Riphean succession in Spitsbergen, perfect in all other respects, is lacking isotopic age determinations except for those rejuvenated in the course of superimposed, predominantly late Caledonian events (Harland *et al.*, 1993). The dates quoted above are estimated on the basis of interregional and intercontinental correlations. The date of 730 Ma for the last excursion is inferred from data on Neoproterozoic rocks in northwestern Canada, where the same negative shift of $\delta^{13}C$ values attributed to tillites of the Rapitan Group in the Mackenzie Mountains and their counterparts in the Victoria Island. These glaciohorizons are correlated with the Sturt tillite in Australia (700–750 Ma). The lower age limit of the Rapitan tillites is defined on the basis of their occurrence above the dikes and diabase sills having the U–Pb zircon age of 766 ± 24 Ma and crosscutting the underlying Little Dal Group (Kaufman and Knoll, 1995). In addition, intrusive bodies having the same position yielded the zircon age of 778–779 Ma, and zircons from granite boulder in tillites turned out to be 755 ± 18 Ma old (Kaufman *et al.*, 1997). The minimum age limit of the glaciohorizon under consideration is established in the Victoria Island, where correlatives of the

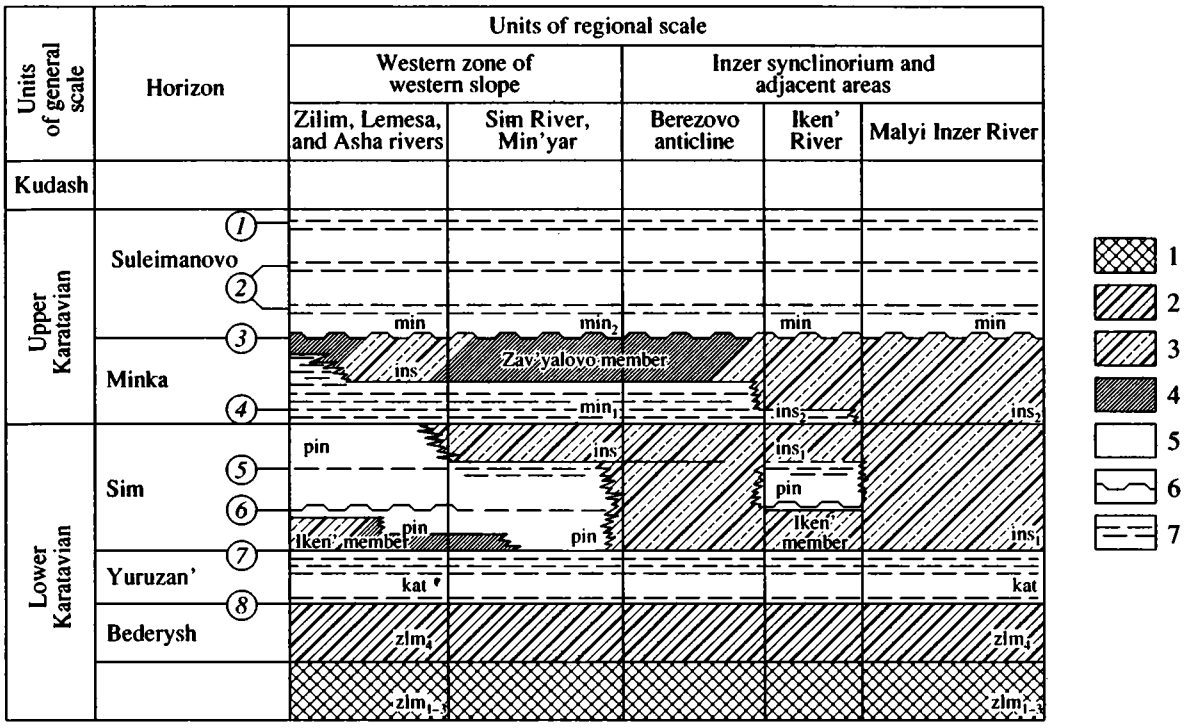

Fig. 4. Alternation of carbonate and siliciclastic deposits in various sections of the Karatau Group, the western slope of southern Urals (after Raaben and Komar, 1983): (1–4) predominantly terrigenous successions, namely, the basal Karatau beds (1), Bederysh Horizon (2), Inzer Formation as cartographic unit (3), and linking units (4) corresponding to Sim and Minka horizons within the Min'yar and Podinzer formations; (5) essentially carbonate successions; (6) beds with *Conophyton*; (7) beds with other organic remains. Encircled numbers denote the following taxa and microphytolith assemblages: *1*, *Heterostylia* associated with III and IV microphytoliths assemblages; *2*, *Minjaria, Parmites nubilosus*, and B'yanka microphytolith assemblage; *3*, *Conophyton miloradovici* var. *krylovi*; *4*, *Gymnosolen ramsayi, Parmites meridionalis*, and Minka microphytolith assemblage; *5*, *Baicalia* and problematics *Paleodyction*; *6*, *Conophyton garganicus* var. *ikeni*; *7*, *Malginella*; *8*, *Inzeria tjomusi*. Abbreviations for stratigraphic subdivisions: (min) Min'yar Formation with Minka (min_1) and carbonate B'yanka (min_2) subformations; (ins) Inzer Formation with lower (ins_1) and upper (ins_2) subformations; (pin) pod-Inzer beds; (kat) Katav formation; (zlm) Zil'merdak Formation with three lower (zlm_{1-3}) and Bederysh (zlm_4) subformations.

Rapitan tillites are overlain by basalts having the U–Pb zircon age of 723 ± 3 Ma (Kaufman and Knoll, 1995, and references therein). In view of listed dates, the time mark of 730 Ma accepted for the considered $\delta^{13}C$ excursion appears to be realistic, although tillites of the same age are unknown in Spitsbergen and Greenland and cannot represent the direct stratigraphic marker for the model curve. Several tillite horizons are also known in Namibia, and one of them that has the U-Pb zircon age of 748 ± 2 Ma was correlated with the discussed glaciohorizon of Canada (Kaufman *et al.*, 1997).

The date of 800 Ma accepted for the preceding negative excursion is tentative without doubts. It is attributed to the U–Pb zircon age of 802 ± 10 Ma measured for the Rook Tuff of the Adelaide "geosyncline" in Australia. According to Kaufman and Knoll (1995, p. 34), if stratigraphic connections between the Adelaide "geosyncline" and Mackenzie Mountains can be proved, this date can define the maximum age of the Shaler carbonates, the basal strata of which correspond in the Arctic Canada to the negative peak under consideration.

The date of 850 Ma accepted for the starting point of the model curve is lacking any concrete isotopic-geochronological substantiation and reflects the changed estimation of the time-range of the Middle Hecla Hoek Supergroup. In the original publication devoted to the C-isotope analysis of this supergroup, it was considered as almost equal in range to the Upper Riphean, and the Veteranen Group was attributed to the first half of the latter (Knoll *et al.*, 1986).

The last viewpoint was compatible with biostratigraphic data on stromatolites (Raaben, 1960, 1975; Golovanov and Raaben, 1967; Golovanov, 1967; Zabrodin, 1968; Raaben and Zabrodin, 1969, 1972; Krasil'shchikov, 1970; Mil'shtein and Golovanov, 1979), and also with conclusions based on the study of microfossils. The monotonous biota characterizing all shally formations of the Veteranen Group was considered to be about 900 Ma old (Knoll and Swett, 1985, 1987) and comparable with biotas from several other sections in various regions, where radiometric dates were available. In particular, it was thought to be similar to biota from rocks of the Upper Riphean type section in the Urals, where they range in age from 940 to 850 Ma (Knoll and Swett, 1985, p. 458). Apparently,

this was reference to the Shishenyak biota described by Yankauskas (1980).

The diverse assemblage of organic-walled and mineralized microfossils appearing in the Svanbergfjellet Formation and characteristic of the upper half of the Academikerbreen Group (Knoll, 1982; Knoll *et al.*, 1986; Butterfield *et al.*, 1994) was attributed to the "younger" Upper Riphean. Its time-range was assumed to be 700–750 Ma on the basis of correlation with the Vinnyat Formation of the Shaler Group, the minimum age limit of which was estimated at that time as equal to 723 ± 3 Ma (see above). This formation yields large acanthomorphic acritarchs of the *Trachyhystrichsphaera/Cymatiosphaera* "Zone," which are typical of the Svanbergfjellet biota.

The changed interpretation of the Veteranen Group age (Kaufman and Knoll, 1995) did not depend on any new biostratigraphic data. Nevertheless, exactly microfossils played an important part in this case. The point is that the Svanbergfjellet Formation turned out to be younger, according to chemostratigraphic data, than the negative peak dated back to 730 Ma, but the Vinnyat Formation is older than that peak. At the same time, Knoll (1994a, 1994b) suggested that the time-ranges of Neoproterozoic assemblages of microfossils are narrow enough to distinguish, in particular, the time interval of 700–750 Ma characterized by the Sveanor biota, and, consequently, he met the necessity to include the Vinnyat Formation in this interval. Considering dating problems further, he had to choose among age values available at that time the minimum one for the lower part of the Shaler Group that was thought to be correlative in Spitsbergen with the lower half of the Academikerbreen Group coupled with the Veteranen Group (Kaufman and Knoll, 1995, p. 41). It should be mentioned here that the refined minimum age limit acceptable for the Shaler Group is within the interval of 778–799 Ma that places the Vinnyat biota outside the time-range mentioned above. In addition, it is even more important that the concept of zonal microfossil assemblages has no support at present. Large acanthomorphic acritarchs *Trachyhystrichosphaera* are most typical of the basal Upper Riphean and exist until the initial Vendian time (Sergeev, 1997). They are present in several early Karatavian biotas, e.g., in the Muldakaevo biota of southern Urals, in the Lakhanda biota of the Yudoma–Maya region, and in the Derevnya biota of the Turukhansk region in Siberia (Yankauskas, 1982; Veis *et al.*, 1998). Almost all taxa of the Sveanor biota are present as well in the Miroedikha biota from the last region of Siberia (Veis *et al.*, 1998), whose age approaches 850–900 Ma, as it is estimated on the basis of paleontological, isotopic-geochronological, and chemostratigraphic data. All information mentioned above does not allow me to agree with Knoll and his colleagues that the stromatolite-based geochronology of the basal Upper Riphean succession in Spitsbergen must be revised.

Similarity of stromatolites from carbonate deposits of Spitsbergen and Karatau Group of the Urals was considered as first indication to the Late Riphean age of the Middle Hecla Hoek Supergroup (Raaben, 1960; Krylov, 1960; Golovanov and Raaben, 1967). The subsequent comprehensive study of stromatolites from Spitsbergen and supplementary data on microphytoliths (Raaben and Zabrodin, 1969, 1972; Krasil'shchikov, 1970) provided grounds to attribute the upper Academikerbreen to the Upper Karatavian and the Veteranen Group to the lower one. In addition, it was established that stromatolite assemblages from the Middle Hecla Hoek Supergroup of Spitsbergen are very similar to those from the Upper Riphean stratotype not only in the taxonomic composition (almost all form genera and nearly a half of form species represent taxa in common for both regions), but also in the succession of stromatolitic beds bearing unique forms and their associations. The lower member in both successions is *Inzeria tjomusi* Kryl. forming biostrome horizons in the lower formation of the Veteranen Group and near the base of the Katav Formation in the stratotype. In the middle part of the Academikerbreen Group and in the Min'yar Formation of the Urals, there are similar successions of beds bearing the following species or their assemblages: (a) *Gymnosolen ramsayi* Steinm. and *Inzeria djejimi* Raab. associated with *Alternella* and *Tungussia* forms endemic in each region; (b) *Conophyton miloradovici* Raab.; and (c) *Minjaria uralica* Kryl. Species *Poludia russa* Raab., the upper member in the succession of common taxa, was originally described as an endemic form of Spitsbergen, but later it was distinguished in the Uk Formation equivalents of northern Urals, and also in the eastern Greenland (Bertrand-Sarfati and Siedlezka, 1980).

The high degree of taxonomic similarity and analogous successions of taxa correlate the Upper Riphean complex of stromatolites from Spitsbergen with their complexes distinguished not only in the southern Urals, but also in many other sections situated along eastern and northern margins of the Upper Riphean carbonate platform of the Russian plate. Elements of that stromatolitic succession, which was established in the Karatau Group, are traceable within the belt extending from the southern and central Urals to the Polyudov Ridge, Timan, Kanin Peninsula, Kildin Island, and further to northern Norway (Raaben and Komar, 1983; Raaben *et al.*, 1995; Raaben, 1997). This situation suggests that all the listed regions may be constituents of a single stromatolitic province despite the peculiar Late Riphean endemics known in Spitsbergen (Golovanov, 1967; Raaben and Zabrodin, 1972; Mil'stein and Golovanov, 1979; Raaben *et al.*, 1995). In general, the succession of stromatolitic beds in Spitsbergen is almost identical to that known in the Urals and central to northern Timan (Raaben and Oparenkova, 1997).

Inside stromatolitic provinces, beds bearing particular stromatolite taxa represent the reliable stratigraphic markers (Semikhatov, 1995). In the case under consideration, these are beds with *Inzeria tjomusi*. As their

stratigraphic position in the stromatolitic succession is stable within a vast territory, we can place with assurance the Veteranen Group yielding this taxon at the Katav level of the Lower Karatavian by analogy with the southern Urals, Polyudov Ridge, and central Timan.

Consequently, the dissimilar configuration of initial segments in the considered C-isotope variation curves is most likely a result of the misleading geochronological interpretation. The comparable succession of large $\delta^{13}C$ anomalies in each curve and the aforementioned biostratigraphic data allow me to suggest that the anomalies established in the Urals and Spitsbergen could be synchronous in both curves. In other words, it is possible to conclude that two positive peaks and separating depression of $\delta^{13}C$ values in the initial segments of the curves, which correspond in the Urals to the Lower Karatavian, should be in phase with the same elements of the model curve, which are attributed at present to the Cryogenian (= Upper Karatavian). As is shown above, this attribution is not substantiated properly, whereas this part of the Uralian curve of C-isotope variations is dated quite confidently. Actually, the Pb–Pb isochron age of lower limestone beds in the Inzer Formation (Podinzer Beds), which limits the interval on top, is 836 ± 24 Ma (Ovchinnikova *et al.*, 1998); K–Ar dates for glauconites from this interval range from 970 to 890 Ma; and remains of *Melanocyrillium* Bloeser, which are typical of deposits younger than 850 Ma (Sergeev, 1997), appear in the Urals at the level of the Min'yar Formation (Podkovyrov *et al.*, 1998, and references therein).

Thus, if the large $\delta^{13}C$ anomalies are correlative in both curves, then the model one spans not only the Cryogenian, but also a part of the Lower Karatavian. On the contrary, if the model curve geochronology is correct, then the anomalies under consideration cannot be regarded as correlative. In the last case, we have to assume that they are missing from the Uralian record because of the incomplete rock succession within the middle of the Karatau Group. Consequently, we should find some structural features in the Min'yar section of the Karatau Group, which could explain this break. As is shown below, it is hardly probable to find evidences of this kind.

Middle segments of curves under discussion reveal one more (third) peak of $\delta^{13}C$ values and adjacent depressions (Fig. 3). In the Min'yar section, this is the interval from the top of the Podinzer Beds (Sim Formation) to the level located a little below the top of the Min'yar Formation According to available isotopic-geochronological data (Podkovyrov *et al.*, 1998), the interval is limited by age values of 836 ± 24 and 720 Ma. In the model curve, the segment with comparable patterns is attributed to the Early Cryogenian bracketed between 800 and 730 Ma. In the reference section of Spitsbergen, it correspond to the lower half of the Academikerbreen Group, namely to the Grusdivbreen Formation and nearby beds of the Svanbergfjellet Formation. In the Urals and Spitsbergen, the correlation of both inphase segments is supported by biostratigraphic data, as the Min'yar stromatolite assemblages appear in those intervals of both successions, which reveal the comparable $\delta^{13}C$ variations. The only difference between them is the lower amplitude of positive $\delta^{13}C$ values in the Uralian curve.

The terminal segment of the model curve corresponds to the upper subdivisions of the Academikerbreen Group, i.e., to the Svanbergfjellet dolomites and to the Draken and Baklundtoppen formations, whereas the termination of the Uralian curve depicts data on upper beds of the Min'yar Formation and on the Uk Formation (see Figs. 1 and 2). As was already mentioned, these segments are dissimilar in configuration. At the same time, they have one feature in common, since the highest $\delta^{13}C$ values recorded in two regions are confined exactly to terminal segments of both curves. The extreme values up to 5.6‰ in the Karatau Group and more than 10‰ in the model region are thought to be specific of the particular Late Riphean time predating the Varanger glaciation (Kaufman and Knoll, 1995).

The available geochronological information is insufficient for evaluation of the precise time-range difference between the rock units demonstrating the peak $\delta^{13}C$ values in the Urals and Spitsbergen, but this inconsistency seems to be quite significant. In the Urals, these values characterize only thin (few meters thick) beds of the Shubino Member crowning the Min'yar Formation. In contrast, the corresponding successions in Spitsbergen, Greenland, northwestern Canada, and Namibia are up to several hundreds meters thick. It is also noteworthy that in Spitsbergen the extreme positive $\delta^{13}C$ values (from 5–6 to 10‰ PDB) alternate in succession with much lower and even negative values (Knoll *et al.*, 1986). Accordingly, the model curve interval in question shows the short-period fluctuations of a considerable amplitude (Figs. 1 and 3). The same frequent oscillations from high positive to low negative $\delta^{13}C$ values are established in the Mackenzie Mountains of northwestern Canada and in Namibia (Kaufman *et al.*, 1997; Hoffman *et al.*, 1998).

The terminal point of the Uralian curve is dated back to 660 Ma on the basis of the Rb–Sr isochron age of glauconite from the base of the Uk Formation, which is equal to 688 ± 20 Ma (Podkovyrov *et al.*, 1998). Taking into account the age of the Varanger glaciation that defines the commencement of the Vendian (Kaufman *et al.*, 1997, and references therein), one should admit that the terminal Late Riphean $\delta^{13}C$ variations recorded in the Mackenzie Mountains and Spitsbergen are missing from the Uralian succession. Moreover, the uppermost member of the Upper Riphean type succession (Krivaya Luka Formation) is lacking carbonate rocks, and this fact indirectly supports the above conclusion. Configuration of the Late Cryogenian segment of the model curve is based on data for upper horizons of the Academikerbreen Group of western Spitsbergen, where

it lacks significant hiatuses and is overlain, without signs of discordance, by the Polarisbreen Group of the Vendian (Harland and Wilson, 1956; Wilson, 1961; Harland *et al.*, 1993; Butterfield *et al.*, 1994).

Since the terminal part of the model curve appears to be representative, I should agree with Podkovyrov and his colleagues (1998) that the peculiar configuration of the same part in the Uralian curve is conditioned by cryptic hiatuses in the corresponding rock succession. The mentioned researchers rightfully rejected the other conceivable factors probably responsible for this dissimilarity.

BREAKS IN THE URALIAN C-ISOTOPE RECORD

The incompleteness of C-isotope record depicted in the Uralian curve reflects breaks in the Upper Karatau carbonate succession, which correspond to siliciclastic members and stratigraphic hiatuses (Podkovyrov *et al.*, 1998). As is assumed, some, probably, considerable intervals are missing from the succession in response to the pre-Vendian, pre-Uk, and pre-Min'yar events, and this should be checked with the help of additional chemostratigraphic investigations (Podkovyrov *et al.*, 1998). These investigations seem to be necessary and may exclude incidental results, which could have origin either in the insufficient sampling density in the laterally variable Karatau Group (Fig. 4), or in the inexact correlation of individual sections integrated into the composite succession (Fig. 2A).

Among those hiatuses, which could be responsible for the stratigraphic incompleteness of the Karatau Group, the pre-Vendian one is most distinct. In the East European platform, Vendian deposits unconformably rest on different subdivisions of the Upper and Lower Precambrian (Bekker, 1988; Aksenov, 1998), and this unconformity is well pronounced in southern Urals. The Vendian Asha Group distinguished here rests on the Uk Formation and overlaps westward different horizons of the Min'yar Formation (*Stratotip...*, 1983; Bekker, 1988). In the Shubino section that was studied in order to depict the terminal segment of the Uralian C-isotope curve, the Uk Formation is transgressively overlapped by the Asha Group (Kozlov, 1982). Even if the Uk Formation would be represented here in its whole range, it should be still separated from the Vendian base by a hiatus corresponding to the accumulation of the siliciclastic Krivaya Luka Formation crowning the Upper Riphean succession in southern Urals. Consequently, the Uralian curve does not reflect the latest Riphean variations of $\delta^{13}C$ values.

The comparison of chemostratigraphic data on the Urals and Spitsbergen also attracts attention to the hiatus predating the Uk Formation. This hiatus is known pretty well (Garan', 1963; Kozlov, 1982; Kozlov *et al.*, 1990, 1997; Shalaginov and Gileva, 1990; Shalaginov and Maslov, 1997), but it used to be considered as insignificant or local because of the suggested structural connections between the Uk Formation and underlying deposits (Keller, 1968; *Stratotip...*, 1983). However, there is a series of geological data suggesting that the missing stratigraphic interval could be rather long. For instance, the boundary between the Min'yar and Uk formations is marked by occurrence of weathered and brecciated Fe–Mn rocks (Domrachev, 1952). At the same level, there are sedimentary iron ore bodies and stratigraphic unconformity, because the Uk Formation normally resting on the terminal, Shubino Member of the Min'yar Formation also overlaps in some places the underlying cherty member of the latter (Bekker, 1961, 1988). In terms of the comparative chemostratigraphy, these geological observations seem to be quite meaningful.

In addition to the considered hiatuses, there is one more possible reason responsible for the incompleteness of the C-isotope record of the Karatau Group (Podkovyrov *et al.*, 1998). I mean shortcomings in correlation of the Min'yar and Shubino sections integrated into the composite succession of the Karatau Group, which was used to reconstruct the Uralian curve of $\delta^{13}C$ variations. The Min'yar Formation sampled near the eponymous town was considered to be 290 m thick (Fig. 2A), but according to all previous data, it is no less than 400–450 m thick there (Bekker, 1961; Krylov, 1963; Raaben, 1975; Raaben and Kozlov, 1983; Maslov, 1988). The accepted value of 290 m almost precisely corresponds to that part of the Min'yar section (265 m), where the formation is exposed continuously, but there is its additional part (150–200 m) exposed fragmentary (Raaben and Kozlov, 1983; Maslov, 1988). It is likely that this additional part (or its greater interval) has not been included into the composite succession. Thus, the rock succession used to depict the Uralian C-isotope variations could be lacking a fragment, which would be able to change terminal patterns of the curve. I hope that further investigations may shed light on the issue, especially if the complete but not composite succession will be studied.

In the Karatau Group interval corresponding to the middle part of $\delta^{13}C$ variation curve, there are two levels with erosion marks. One is at the top of the siliciclastic Zav'yalovo Member crowning the upper Inzer Subformation and another near the base of the Minka Beds. Both levels reveal signs of subaerial exposure that gave rise to Fe-oxide concentration and to deposition of insignificant ore bodies (Bekker, 1961; Raaben, 1975, 1981). However, they are unimportant for consideration, as none of them is detectable in the comparative chemostratigraphic plot. Similarly insignificant is the siliciclastic member underlying the Minka Beds (Podkovyrov *et al.*, 1998).

The interval of the Podinzer Beds is of particular interest, because it corresponds to the chronometric interval of 800–850 Ma, i.e., to the model curve segment with three high-amplitude anomalies of $\delta^{13}C$ values. The Pb–Pb isochron age of 836 ± 24 Ma obtained for the Podinzer Beds (Ovchinnikova *et al.*, 1998) is most reliable among other dates characterizing the

Karatau Group and coincides, within the error limits, with the chronometric interval mentioned above.

Data of regional geology exclude any chance that the succession of Podinzer Beds (Sim Formation) may be interrupted by hiatuses. Near the town of Min'yar, the base of this carbonate beds overlying a thin (5–7 m) silty–pelitic member is also lacking erosion marks. Hiatuses are also missing from those western sections of the beds, where the silty–pelitic unit is distinguished as the particular Iken' Member owing to its greater thickness (Raaben and Kozlov, 1983, and references therein). In eastern sections, where the Inzer Formation is entirely siliciclastic, transition from the underlying carbonates of the Katav Formation to these siliciclastic deposits is gradual, as it is described in many publications, beginning from the early book by Olli (1948) and ending with most recent summarizing works. This transition was several times studied in detail, because it was thought to be either indicative of the transgressive overlap between the Inzer Formation and underlying Podinzer and Katav beds (L'vov and Kondiain, 1956), or dividing two cycles of the Urals evolution during the Late Riphean (Olli, 1948; Bekker, 1972, 1994). As a result, the absence of hiatuses within lower horizons of the Karatau Group is well established. Consequently, the dissimilar configuration of the model and Uralian C-isotope curves in their initial parts cannot be related to the incompleteness of the Karatau Group succession. This discordance is merely apparent, related to misleading age constraints for the corresponding part of the model curve, which were accepted without proper attention to comparative biostratigraphic data on the Upper Riphean stromatolites from the southern Urals and Spitsbergen.

CONCLUSION

The C-isotope variation curve, which approximates data on the Upper Riphean deposits of Spitsbergen, Greenland, northwestern Canada, and southwestern Africa (Kaufman and Knoll, 1995), is now accepted as the model one for the second half of this stratigraphic interval and as useful for global correlations. The careful analysis of the used chronometric calibration shows that it is vulnerable in the initial, relatively older part of the curve. Nevertheless, the model curve and another one recently obtained for the Karatau Group, the Upper Riphean stratotype in southern Urals (Podkovyrov *et al.*, 1998) are of the in-phase character. They both demonstrate the identical number of large positive and negative peaks despite the lower amplitudes of $\delta^{13}C$ variations in the Uralian section that is conditioned, in general, by lower values of the extreme positive points. The in-phase character of both curves along with biostratigraphic, isotopic-geochronological and regional geological data suggest that the so-called model curve characterizes not only the second half of the Late Riphean, but nearly its entire interval. The observable difference in configuration of the uppermost parts of the curves under comparison appears to be a consequence of the Upper Riphean stratotype incompleteness, as there is a large pre-Vendian stratigraphic hiatus and another one below the Uk Formation, which was underestimated in previous works. Shortcomings in chronometric calibration of the model curve and newly obtained data on C-isotope variations in the Karatau Group of the Urals indicated that many chemostratigraphic correlations between the Upper Riphean successions should be revised.

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On Possibility of Using the Global Permian Stage Scale in the Tethyan Region

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Abstract—A possibility of using the Permian global chronostratigraphic scale in the Tethyan region is estimated in this work. Permian successions in this region show that the proposed subdivision of the Permian System into the Cisuralian, Guadalupian, and Lopingian series is efficient when it is definitely proved that the lower boundary of the Guadalupian Series corresponds to the base of the Kubergandian Stage of the Tethyan scale. This problem is, however, unsolved so far. The Asselian, Sakmarian, and Artinskian stages of the Cisuralian Series correspond to the Asselian, Sakmarian, and Yakhtashian stages of the Tethyan scale. The correlation between the Kungurian and Bolorian stages can be taken for testified, when the exact position of the lower boundary of the Guadalupian Series is established. The conodont biostratigraphy used to discriminate the Roadian, Wordian, and Capitanian stages of the Guadalupian Series is inappropriate for the Tethyan region. The Wuchapingian and Changhsingian stages of the Lopingian Series correspond to the Dzhulfian (sensu stricto) and Dorashamian stages of the Tethyan scale.

Key words: global chronostratigraphic scale, Tethyan scale, correlation, Cisuralian, Guadalupian, and Lopingian series, Asselian, Sakmarian, Artinskian, Kungurian, Roadian, Wordian, Capitanian, Wuchapingian, Changhsingian, Yakhtashian, Bolorian, Kubergandian, Murgabian, and Midian stages.

INTRODUCTION

The recent activity of the International Subcommission on the Permian Stratigraphy (SPS) resulted in elaboration of the global chronostratigraphic scale of the Permian System (Jin *et al.*, 1997). In the latter, the system is subdivided into three series: Cisuralian, Guadalupian, and Lopingian. The first of them corresponds to the lower series of the traditional European scale, the second one is defined in sequences of the Guadalupe Mountains in western Texas, and the third unit is distinguished in South Chinese sections. The stage subdivision of series is carried out in line with their subdivision in the relevant type regions. In December 1999, the SPS approved by voting the Guadalupian Series and its constituent Roadian, Wordian, and Capitanian stages. This decision was ratified by the International Geological Congress held in Rio de Janeiro in August 2000. The Cisuralian and Lopingian series are next to be officially approved. Thus, in the nearest future, one should expect an official approval of the new stage scale of the entire Permian System. This scale will unavoidably create many problems and difficulties in applying the new criteria for positioning the stage boundaries and their tracing beyond the location areas of the type sections.

This paper is aimed to assess the new scale from the standpoint of its application in the Tethyan region and to consider the relevant problems and possible ways of their solution.

STANDARD AND TETHYAN STAGE SCALES AND PRINCIPLES OF THEIR ELABORATION

Miklukho-Maklay (1958) was the first to propose the initial version of the stage scale for the Caucasian–Sinian biogeographic province. Disclaiming a possibility to apply the East European stage scale for this region, he defined, subdividing the system into two series, four new stages: the Karachatyrian and Darvazian in the lower series, and the Murgabian and Pamirian in the upper one. After subsequent works (Leven, 1963, 1967, 1974, 1975, 1979, 1980a, 1980b, 1981; Rostovtsev and Azaryan, 1974), the scale was significantly modified and detailed. In 1979, the Interdepartmental Stratigraphic Committee (ISC) of the former Soviet Union recommended it for using in the southern (Tethyan) regions of the USSR. The version approved by the ISC was published in 1980 (Leven, 1980a) and, since that time, it was widely used by Russian and foreign geologists as a standard for subdivision and correlation of Permian deposits in various areas of the Tethyan region. In the published version, the Permian System was subdivided, in line with tradition, into two series. It should be mentioned, however, that previously I proposed a three-member scheme for the Permian System respectively subdivided into the Yaikan, Kushanian, and Arianian series (Leven, 1974, 1975). A possibility of subdividing the Permian System into four series constituting two superseries was discussed as well (Leven, 1992a, 1992b, 1997; Kotlyar *et al.*, 1994).

All stages of the Tethyan scale, except for two upper ones, were discriminated on the basis of fusulinids. The Asselian and Sakmarian stages were established by means of correlation of their characteristic fusulinid assemblages from the Uralian type sections with the counterparts from the Tethyan successions. Most of remaining stages were recognized on the basis of natural phases in evolution of higher fusulinids. The stage boundaries were placed at the first occurrence levels of taxa of the generic and family ranks in their continuous phylogenetic lineages, beginning from primitive *Pamirina* to highly organized *Yabeina* and *Lepidolina* forms. Changes in the accompanying fusulinid assemblages were also taken into consideration. Such an approach to definition of stages ensured the continuity of their succession and unambiguous understanding of boundaries between them. Along with a wide geographical distribution and frequent occurrence of fusulinids, this enabled a relatively easy recognition of stages in the entire Tethyan region from the Mediterranean to Japan and Indochina. At the end of the Midian Age, most of fusulinids became extinct, and, as a consequence, two upper stages—Dzhulfian and Dorashamian—were established using the ammonoid biostratigraphy, and later were supported by conodont successions.

As was shown in several works (Leven, 1993; Leven *et al.*, 1996), the stages in fusulinid development were related to major abiotic events of trans-Tethyan extent and, first of all, sea-level oscillations. The latter affected the entire biota, and the main turning points in fusulinid development and the relevant boundaries between stages and series were substantiated, to a different extent, with due consideration of other faunal groups, e.g., of ammonoids, brachiopods, and conodonts.

As is mentioned in the introduction, the global chronostratigraphic scale of the Permian System is composite. Its three series are of the natural–historical character, reflecting major stages in the marine biota development. Stages of the series are established by representatives of different geological schools in different periods and in various regions of different biogeographic provinces. In contrast to the Tethyan scale, they are defined, therefore, on the basis of different principles. For instance, the Asselian, Sakmarian, and Artinskian stages established in the Urals correspond to certain stages in ammonoid evolution (Karpinsky, 1874; Ruzhentsev, 1951, 1954, 1956). Boundaries between these stages were specified according to fusulinid biostratigraphy. The Kungurian Stage was originally established as a local lithostratigraphic unit (Schtuckenberg, 1890), and its biostratigraphic characteristic is incompletely clear even now. The Roadian, Wordian, and Capitanian stages of the Guadalupian Series initially represented the synonymous formations; the stage status was attributed to them later, after elucidation of their paleontological characteristics. In the most clear and logical form, this was done by Furnish (1973). Defining the Roadian Stage as the terminal one in the Lower Permian, he indicated its characteristic assemblage of ammonoids, most typical among which are *Perrinites hilli* (Smith), *Paraceltites elegans* Girty, *Texoceras texanum* (Girty), *Glassoceras normani* (Miller et Furnish), and some other forms. Furnish correlated the Wordian Stage with the *Waagenoceras* Zone, and the Capitanian Stage with the *Timorites* Zone. In addition, the Amarassian Stage characterizing the first occurrence of *Cyclolobus* forms was defined in the Guadalupian Series. In general, data on fusulinids are consistent with his conclusions, although they did not reveal such a distinct picture. For instance, in the opinion of Ross (1963), the Roadian fusulinid assemblage is almost identical to the Wordian one. The boundary between the Wordian and Capitanian stages is marked by the first occurrence of *Polydiexodina* forms. The fusulinid assemblage from the Lamar Limestones referred by Furnish to the Amarassian Stage also bears characteristic features. However, no ammonoids have been found in the limestones, and their correspondence to the stage under consideration is not unambiguous.

Because of a wide distribution of ammonoids and quite distinct differences between their assemblages, the stage scale suggested by Furnish for the Guadalupian Series was accepted worldwide. However, when this series and its stages were proposed to be the international standard units, it appeared that ammonoids are unsuitable for defining the Global Stratotype Sections and Points (GSSP) because they occur only episodically in the type sections. It was impossible also to use fusulinids and brachiopods for this purpose because of their distinct endemism. As a result, conodonts turned out to be the only group appropriate for substantiation of the Guadalupian stages and their boundaries. In the study of conodonts, the main attention was paid to the search of their continuous phylogenetic lineages, which would allow to trace species succession and, thus, to tie the GSSP to the particular moments of their evolutionary changes without the necessary analysis of this faunal group evolution. The procedure of discriminating the stages was formalized and eventually reduced to hammering the "gold spikes" without the proper consideration of successive geological events and general development of biota. In this new interpretation, substantially different time-ranges were attributed to the stages named in accordance with the nomenclature by Furnish, and his criteria lost their significance to a considerable extent. In that new interpretation, stages correspond to the of conodont *Jinogondolella nankingensis* (Roadian Stage), *J. aserrata* (Wordian Stage), and *J. postserrata* (Capitanian Stage) phylogenetic zones distinguished in the sequence. Boundaries between stages are placed at the first occurrence levels of zonal species, which are recorded approximately in the middle of synonymous formations and above the levels marked by renewals in ammonoid and fusulinid assemblages (Fig. 1).

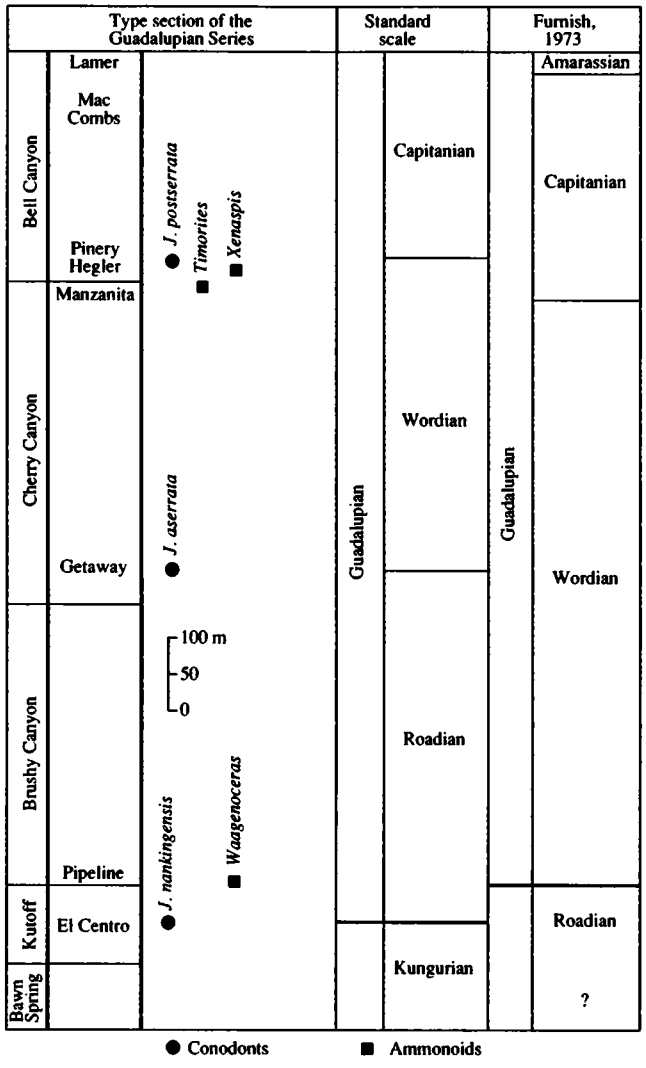

Fig. 1. Type section of the Guadalupian Series and variants of its subdivision into stages separated by the first occurrence levels of conodonts (standard scale) and ammonoids (scheme by Furnish).

Such an exclusive attention to conodonts, when the comprehensive definition of stages and their boundaries was actually discarded, limited the possibility of tracing these subdivisions and, consequently, of using them as global standards. The declared cosmopolitanism of conodonts is obviously overestimated. They are rare in the Boreal and Notal provinces, and less common than fusulinids in the Tethyan region. Similarly to other faunal groups, distribution of conodonts depends on environments and facies (Kozur, 1995, 1998). All this does not allow me to share optimism of those who thinks that the Guadalupian stages can be used beyond the Midcontinent and should be included into the global scale.

Subdivisions of the Lopingian Series of the standard scale, i.e., the Wuchapingian and Changhsingian stages, are established in the South Chinese sequences. Their paleontological characteristics are good and versatile (ammonoids, fusulinids, conodonts), and their boundaries are defined on the basis of different indicators. All these indicators are workable, however, in the Tethyan region only, and both stages can hardly be recognized in other biogeographic provinces in the nearest future.

All the aforesaid shows that the proposed version of a new global chronostratigraphic scale is inappropriate for solving any of the problems of the Permian stratigraphy and, to the contrary, it complicates them sometimes. First of all, this is a consequence of inadequate substantiation of the Guadalupian stages and their boundaries, which can be hardly recognized beyond the stratotype region. A possibility of using the new Permian global scale in the Tethyan region is evaluated below.

BOUNDARIES OF THE PERMIAN SYSTEM

The lower Permian boundary in the global chronostratigraphic scale is established in the Aidaralash section in accordance with the conodont biostratigraphy (Davydov *et al.*, 1995). Its position is also compatible with distribution patterns of ammonoids and fusulinids, and the boundary is easily recognizable in many sections throughout the Tethyan region owing to the successive fusulinid assemblages of the *Ultradaixina bosbytauensis* and *Sphaeroschwagerina vulgaris–Sphaeroschwagerina fusiformis* zones (Leven and Davydov, 1979; Watanabe, 1991; Leven, 1992b; Davydov, 1995; Krainer and Davydov, 1998).

The upper boundary of the Permian System is not officially approved so far. In the global scale, it is placed at the base of the conodont *Hindeodus parvus* Zone, as it was recommended by the SPS Working Group on the Permian–Triassic boundary (Yin, 1997; *Vote on...*, 1999). Inasmuch as these recommendations are based on biostratigraphic investigations in South China, we have no problem with tracing this level throughout the Tethyan region. The boundary is also recognizable in some Boreal sections, where its position is established within the upper part of the *Otoceras boreale* Zone of the ammonoid scale (Kozur, 1998).

SERIES OF THE PERMIAN SYSTEM

In the global scale, the Permian System includes three series. In principle, this is quite acceptable for the Tethys, because many experts on the Permian stratigraphy in this region used the three-member scale for a long time. Indeed, the Middle Permian begins from the level corresponding to a large-scale Kubergandian transgression that was responsible for a significant reorganization of the entire marine biota (Leven, 1993; Leven *et al.*, 1996). For a long time, Chinese geologists correlated the Carboniferous–Permian boundary with this level. The Middle–Late Permian boundary marks the catastrophic extinction of most groups of marine

organisms. This event was probably related to a rapid regression and signified the commencement of the Great Late Permian extinction (Leven, 1993; Leven and Korchagin, in press; Jin, 1993; Jin *et al.*, 1994). Thus, the idea of three-member subdivision of the Permian System is not new and sufficiently consistent with natural events in the Permian history. Therefore, the Guadalupian Series incorporation into the global scale should be welcomed. It should be emphasized, however, that the first occurrence level of the conodont species *Jinogondolella nankingensis* used to define the lower boundary of the series in the type sections of the North American Delaware basin cannot be effectively applied for recognizing this boundary in the Tethyan region, where the species in question is extremely rare. This question is discussed in the next section, where the correlation of stages and their boundaries is under consideration.

The upper boundary of the middle series that is established in sections of South China at the top of the Maokou Limestones (Jin, 1999) corresponds to the top of the Capitan Formation in the Delaware basin, where this level is well recognizable on the basis of conodont zonation (Mei *et al.*, 1998).

STAGES OF THE PERMIAN SYSTEM

Asselian Stage. All stages included into the lower Cisuralian Series of the Permian System were established in the Urals. The possibility of practicing them in the Tethyan region was repeatedly discussed (Leven, 1967, 1979, 1980b; Leven and Shcherbovich, 1978; Leven and Davydov, 1991; Leven *et al.*, 1992; Movshovich *et al.*, 1979; and others). It became clear after the discussion that recognition of the Uralian stages in the Tethyan region is not an easy task because of isolation of the East European basin that commenced in the Sakmarian Age. This general conclusion remains valid until present. The free water exchange that existed between the Uralian and Tethyan basins in the earliest Permian time favored the faunal interchange, and, correspondingly, the Asselian Stage is easily recognizable on the basis fusulinid biostratigraphy throughout the entire Tethyan region. Similarly to the South Uralian stratotype region, the Asselian fusulinid assemblage from the Tethys includes subspherical forms with a freely coiled spire, which belong to genera *Sphaeroschwagerina, Pseudoschwagerina, Paraschwagerina, Zellia* and coexist with abundant *Rugosofusulina, Dutkevitchia*, and *Pseudofusulina* species. Particularly close to the South Uralian faunas are fusulinid assemblages from adjacent areas of the Tethyan region, e.g., from the Fergana and Darvaz. Accordingly, the Asselian Stage in these areas can be subdivided into three fusulinid zones recognized in the Urals (Bensh, 1972; Leven and Shcherbovich, 1978; Leven and Davydov, 1991). The Asselian fusulinids are missing, however, from the southern peri-Gondwanan areas of the Tethys (southern Afghanistan, southern Pamirs, Karakorum, Himalayas, and southern Tibet), and it is impossible so far to recognize the Asselian Stage there. All attempts to distinguish this unit on the basis of other faunal groups seem doubtful, because they usually operate with indirect data instead of direct correlation with the type sections of the stratotype region.

Sakmarian Stage. Like the Asselian Stage, the Sakmarian one is also included into the Tethyan scale, although its recognition in the region under consideration meets more difficulties than in the previous case. Two reasons are responsible for this: the definition of the stage boundaries in the stratotype region is inexact, and the isolation of East European basins from the Tethys commenced at that time.

When establishing the Sakmarian Stage and characterizing ammonoid assemblages of the stage as a whole and of the constituent Tastuba and Sterlitamak horizons, V.E. Ruzhentsev put aside the problem of its lower boundary, because sections displaying the succession of these assemblages are unknown. Placing this boundary at the top of the Kurmaya Formation in the Kondurovka section of the Orenburg region of the Urals, he paid main attention to fusulinids, but not to ammonoids. According to Rauser-Chernousova (1965), the boundary in question marks the major reorganization in fusulinid communities. The Asselian assemblage with characteristic *Sphaeroschwagerina, Pseudoschwagerina, Rugososchwagerina*, and *Dutkevitchia* forms is replaced here by the impoverished assemblage mainly consisting of *Pseudofusulina* species. Most remarkable among the latter is *Pseudofusulina moelleri* (Schellwien) from the lower zone of the Sakmarian Stage. Characteristic of this zone are also paraschwagerins, for instance, *Paraschwagerina mira* Rauser.

In the Tethyan sequences, the Sakmarian Stage is recognized on the basis of fusulinids despite a substantial difference between their assemblage from this and stratotype regions. For instance, the Sakmarian Stage hosts here almost all genera and many species characteristic of the Asselian Stage. Accordingly, the boundary between these stages is obscure in the Tethyan region. Nonetheless, the appearance of abundant paraschwagerins, such as *P. mira*, and of pseudofusulins from the *P. moelleri* group elucidates more or less confidently the boundary position (Bensh, 1972; Leven and Shcherbovich, 1980). In sections of the Darvaz, Japan, and Carnic Alps, this boundary corresponds to the first occurrence level of the *Robustoschwagerina* and *Zellia* forms, e.g., of *R. schwellwieni* (Hanzawa), *R. geyeri* (F. et G. Kahler), *Z. heritschi* F. et G. Kahler, and also of *Darvasites* species (Leven and Shcherbovich, 1978, 1980; Watanabe, 1991; Forke, 1995). Beds yielding all mentioned fusulinid genera and species in association with abundant representatives of *Pseudofusulina* and *Rugosofusulina* genera are considered as corresponding to the *Robustoschwagerina–Paraschwagerina* Genus-zone referred to the Sakmarian Stage, although one cannot be completely sure that boundaries of this

zone (particularly the upper one) exactly coincide with those of the stage.

The inclusion of the *Robustoschwagerina–Paraschwagerina* Zone into the Sakmarian Stage appears to be correct also from data on distribution of ammonoids and conodonts. For instance, in the Darvaz sections, the fusulinid assemblage of this zone occurs together with ammonoids, among which several species of the *Boesites, Synartinskia, Medlicottia*, and *Paragastrioceras* genera are similar to the species from Sakmarian deposits of the Uralian region. In opinion of Leonova, this assemblage shows that enclosing rocks are correlative with the Sakmarian Stage (Leven *et al.*, 1992). In the Carnic Alps, similar fusulinids are found together with conodonts *Mesogondolella* cf. *bisselli* (Clark et Behnken), *Sweetognathus inornatus* Ritter, and *S.* aff. *whitei* (Rhodes) characteristic of the uppermost Sakmarian and lowermost Artinskian stages (Forke, 1995). Similar fossils are recorded in sections of Japan and China.

Artinskian Stage. In the Tethyan scale, the Artinskian Stage corresponds to the Yakhtashian one. According to fusulinid biostratigraphy, the latter is equivalent to the *Pamirina–Chalaroschwagerina* Genus-zone. Its lower boundary coincides with the first occurrence level of representatives of these genera, and the upper one is marked by the first occurrence of the genus *Misellina* represented by the subgenus *Brevaxina*. In contrast to the Sakmarian fusulinid assemblage from the Tethyan and East European basins, which show some similarity, the coeval Artinskian and Yakhtashian assemblages are different absolutely and cannot be used for the direct correlation of the stages. The correlation is possible only on the basis of ammonoids and conodonts.

For instance, the upper part of the Yakhtashian Stage stratotype yields a diverse ammonoid assemblage associated with characteristic fusulinids *Pamirina darvasica* Leven, *Chalaroschwagerina vulgaris* (Schwellwien et Dyhrenfurth), *Darvasites ordinatus* (Chen), *Pseudofusulina kraffti* (Schwellwien et Dyhrenfurth). According to Leonova, the ammonoid assemblage includes species of the *Daraelites, Propinacoceras, Thalassoceras, Metalegoceras, Eothinites, Kargalites, Almites, Cardiella, Crimites*, and *Neocrimites* genera, which are "... close to the Uralian late Artinskian species of the corresponding genera" in the development level. This suggests that "... the upper part of the Yakhtashian Stage can be correlated with the upper half of the Artinskian Stage of the Urals" (Leven *et al.*, 1992). Conodonts found together with ammonoids are represented by *Mesogondolella bisselli* (Clarke et Behnken) and *Sweetognathus inornatus* Ritter, all occurring in the uppermost Sakmarian and Artinskian stages of the Uralian sections. Similar data are known from the South China, where the *Pamirina darvasica* Beds (Longyin Formation or Longyinian Stage) yield also the Artinskian ammonoids *Propinacoceras simile* Haniel, *Popanoceras kueichowense* (Chao), *Daraelites* cf. *elegans* Tchernow, *Eothinites* sp., and others, as well as conodonts *Sweetognathus whitei* (Rhodes) (Zhou, 1988, 1989a, 1989b; Sheng and Jin, 1994).

The cited data indicate that the Yakhtashian Stage corresponds approximately to the Artinskian one. Nonetheless, only more exact correlation of their boundaries, position of which should be specified with the help of additional sampling of ammonoids and conodonts in the Uralian and Tethyan sections, may prove the complete correspondence of these units. In addition, the improved correlation of the conodont, ammonoid, and fusulinid zonal scales is necessary.

Kungurian Stage. This unit is considered as equivalent to the Bolorian Stage of the Tethyan scale, because the latter occurs above the Yakhtashian Stage, which is, in turn, correlated with the Artinskian one. The direct correlation of the Bolorian and Kungurian stages is hardly possible due to the poor paleontological characteristics of the latter in the Urals, where its characteristic fossils are mostly endemics. Extremely rare fusulinids of the Kungurian Stage are represented by *Parafusulina* species similar to those from the uppermost Artinskian deposits and missing from the Tethyan sections. Ammonoids from the lower part of the stage (the Sarana Horizon) are similar to those from the uppermost Artinskian Stage. Upward in the section, their assemblage becomes impoverished and, already in the middle part of the stage, ammonoids are almost exclusively represented by endemic species of the *Uraloceras* genus inherited from the Artinskian Stage. The diverse Buztera ammonoid assemblage is recorded in sections of southeastern Pamirs within beds with typical Bolorian fusulinids. In opinion of Leonova (Leonova and Dmitriev, 1989), its constituting forms are "... slightly younger as compared with species of the Baigendzha (upper Artinskian) assemblage from the Urals, and the corresponding Bolorian deposits occupy position in the global stratigraphic scale above the Artinskian Stage and correspond to the Kungurian level." Analyzing distribution of conodonts, Kozur arrived at a similar conclusion (Kozur, 1995, 1998; Kozur *et al.*, 1994). In the southeastern Pamirs, he distinguished two conodont assemblages characterizing the considered interval of the Permian sequence: *Mesogondolella shindyensis–M. intermedia* and *Neostreptognathodus idahoensis–N. leonovae*. The first of them includes *Neostreptognathodus exculptus* Igo, a species that appears in association with *N. pnevi* Kozur et Movshovich in the western areas of North America. The latter is, in turn, characteristic of basal layers of the Sarana Horizon in the Uralian sections (Sylva Reefs and Shurtan Beds), and its first occurrence level is now accepted for the lower boundary of the Kungurian Stage (Kozur, 1995, 1996, 1998; Jin *et al.*, 1997; Chuvashov *et al.*, 1999). *N. idahoensis* (Youngquist, Hawley et Miller) from the second assemblage is a species occurring worldwide in the Kungurian and coeval deposits.

Thus, the cited data suggest an approximate correlation of the Kungurian and Bolorian stages, although, as in the case with the Sakmarian and Artinskian stages, the exact correlation between boundaries of these units should be verified. Indeed, there is no evidence indicating that the first appearance of *Misellina* forms marking the base of the Bolorian Stage was synchronous to first occurrence of *N. exculptus* and *N. pnevi* at the lower boundary of the Kungurian Stage. The degree of correspondence between upper boundaries of these stages will be discussed below.

Roadian Stage. As was mentioned, this unit was established by Furnish (1973) on the basis of ammonoids. The stage assemblage includes first representatives of the Ceratitida order (genus *Paraceltites*), new *Glassoceras* and *Texoceras* genera, and new species of the *Perrinites, Eumedlicottia*, and *Peritrochia* genera discovered in the Road Canyon Formation of the Glass Mountains in Texas. A similar assemblage is also recorded in the Kutoff Formation of the Guadalupe Mountains and in Meade Peak Beds of the Phosphoria Formation in Idaho. The assemblage from the latter locality includes also the *Daubichites* genus widespread in many sections of the Boreal province, South China and, even, in Australia. This fact seems to be opportune for the global recognition of the Roadian Stage, but it is not, as is shown below.

Furnish did not define the exact boundaries of the Roadian Stage and its stratotype. His data and materials of other specialists on ammonoids, who accept the Roadian Stage as valid, do not answer the question how quickly the Cathedral (Artinskian, according to Furnish, and Kungurian, according to the modern view) assemblage was replaced by the Roadian one, and to what extent concurrent are the appearance levels of its genera and species. Nothing exact was also said about the upper boundary of the stage, although one can understand that it was correlated with the first occurrence level of the *Waagenoceras* genus, which is now universally accepted for that boundary.

In the global stage scale under discussion, the problem concerning the stratigraphic range and boundaries of the Roadian Stage is solved with the help of conodonts (Glenister *et al.*, 1999). The time-range of the stage is correlated with that of the *Jinogondolella nankingensis* conodont zone. Its lower boundary is placed at the first occurrence level of the zonal species representing a descendant of *Mesogondolella idahoensis idahoensis* (Youngquist, Hawley et Miller). In the Guadalupe Mountains of Texas, this event is recorded in the upper part of the El Centro Bed, 43 m above the base of the Kutoff Formation (Fig. 1). The GSSP of the lower boundary of the stage and, correspondingly, of the entire Guadalupian Group is fixed precisely at this level. In the Glass Mountains, the replacement of *M. idahoensis idahoensis* (Youngquist, Hawley et Miller) by *J. nankingensis* (Ching) occurs in the lower third of the Road Formation (Wardlaw and Grant, 1987). The Roadian–Wordian boundary is placed at the first occurrence level of *Jinogondolella aserrata* (Clark et Behnken), the descendant of *J. nankingensis*. The boundary GSSP is located within the Getaway Limestones of the Cherry Canyon Formation of the Guadalupe Mountains. In the Glass Mountains, the boundary is drawn between the Willis Ranch and Appel Ranch beds of the Word Formation, if we consider classical definition of the latter. In both cases, it occupies a substantially higher position than the first occurrence level of the ammonoid genus *Waagenoceras*. Correspondingly, the time-range of the Roadian Stage in the new (conodont-based) interpretation exceeds that proposed by Furnish on the basis of ammonoids.

Fusulinid assemblages of the Roadian Stage are uniform, mainly represented by species of the *Parafusulina* genus. Boundaries of the stage inferable from fusulinid biostratigraphy are very ambiguous. The endemic species composition of Roadian parafusulins is unfavorable for distant correlations, with the Tethyan sections in particular. In the last case, conodonts and ammonoids are of the prime significance. As is shown below, the Roadian Stage is now traced from the stratotype region (the Delaware basin of North America) to the Tethys. In the Tethyan sections, the ammonoid assemblage comparable with the Roadian one is discovered in the southeastern Pamirs, Afghanistan, presumably, in the Crimea, and also in some sections of South China.

In the first region, this is the so-called "Kubergandian assemblage" described by Tumanskaya (1935) who correlated it with the Wordian assemblage of Texas. It should be noted that beds subsequently attributed to the Roadian unit were included at that time into the Word Formation. Bogoslovskaya was first to correlate the Kubergandian assemblage with the Roadian one. In her opinion, "the Kubergandian ammonoids are undoubtedly younger than those of the Artinskian–Kungurian level, because they include species of the *Epiglyphioceras, Stacheoceras*, and *Paraceltites* genera that appear first in the Roadian assemblage of North America... Simultaneously, the Kubergandian Ammonoidea are definitely older than their Wordian assemblage, because the *Bamyaniceras* and *Popanoceras* genera unknown above the Roadian are present among them..." (Chediya *et al.*, 1986). In the stratotype section (the Kubergandy River, southeastern Pamirs), ammonoids are confined to the lower *Armenina–Misellina ovalis* Zone of the Kubergandian Stage. Occurring directly lower in the section are deposits of the Bolorian Stage with *Misellina termieri* (Deprat), *M. parvicostata* (Deprat), *Brevaxina otakiensis* (Fujimoto), and others. Associated forms are *Misellina ovalis* (Deprat), *Armenina salgirica* A. M.-Maklay, first *Yangchienia* and *Pseudodoliolina* (Leven, 1981; Chediya *et al.*, 1986).

In the Pamirs, the upper *Cancellina cutalensis* Zone of the Kubergandian Stage is barren of ammonoids, but

the latter are well represented in this interval of the section in the Tezak exposure of Central Afghanistan (Termier *et al.*, 1972; Desparmet *et al.*, 1972). The corresponding ammonoid assemblage is almost identical to the Kubergandian one in the generic and, to a large extent, in the species composition and can be considered, therefore, as correlative with the Roadian assemblage. Ammonoids associate with fusulinids of the *Cancellina cutalensis* Zone (Leven, 1997). Slightly higher in the section, there are fusulinids typical of the lowermost Murgabian strata, such as *Neoschwagerina simplex* Ozawa and *Presumatrina neoschwagerinoides* (Deprat).

Ammonoids coexisting with fusulinids of the lower zone of the Murgabian Stage were also found in Sample 110/1 from the Kichkhi-Burnu outcrop at the Marta River in the Crimea (Kotlyar *et al.*, 1999). At this site, the assemblage includes *Propinacoceras* sp., *Prostacheoceras tauricum* (Toumansky), and *Cardiella (?) kussica* (Toumansky). Zakharov who determined these fossils considers the assemblage as the Kubergandian or Roadian in age, but his conclusion is insufficiently substantiated in opinion of Bogoslovskaya (oral communication). In addition to his own materials, Zakharov cited also a large list of taxa described by Tumanskaya (1931) from the Kichkhi-Burnu site (the Marta assemblage). The section exposed at this site consists, however, of several limestone blocks, which yield, along with the early Murgabian species, also fusulinids from higher zones of this stage (Miklukho-Maklay, 1957). Accordingly, it is not inconceivable that ammonoids were sampled by Tumanskaya precisely from these blocks and should be younger in age. Tumanskaya correlated them with the Sicilian (i.e., Wordian) forms, and Bogoslovskaya (1984) supported this conclusion.

In South China, the ammonoid assemblage from the Kufeng Formation that includes the *Kufengoceras, Altudoceras, Paraceltites, Daubichites*, and *Cibolites* genera is usually correlated with the Roadian one (Zhou, 1986, 1989a; Sheng and Jin, 1994). The listed ammonoids do not associate with fusulinids, and their exact position in the Tethyan scale is unclear. It is inferred from the fact that the Kufeng Formation overlies the Chihsia Formation and replaces the Maokou Formation or some its beds along the strike. In recent works, Chinese stratigraphers defined in the Kufeng Stage two *Neoschwagerina craticulifera* and *Neoschwagerina margaritae* fusulinid zones corresponding to the upper half of the Murgabian Stage, or three *Altudoceras–Paraceltites, Kufengoceras*, and *Waagenoceras* cephalopod zones (Sheng and Jin, 1994). Ammonoids from two lower zones are correlated with the Roadian assemblage.

Summarizing these data on distribution of ammonoids of the Roadian type in the Tethyan sections, one can find out that they occur from the base of the Kubergandian Stage (the *Armenina–Misellina ovalis* Zone) to the middle part of the Murgabian Stage (the *Neoschwagerina craticulifera*) at least. According to ammonoid biostratigraphy, the lower boundary of the Roadian Stage and, correspondingly, of the Guadalupian Series, corresponds to the base of the Kubergandian Stage. In the Tethyan sections, this level is marked by significant changes in composition of fusulinid and cephalopod communities and, thus, it is most suitable for placing here the boundary between series of the Permian System (Leven *et al.*, 1996). As for the upper boundary of the Roadian Stage in the Tethyan sections, it is not determined using ammonoids because of two reasons. First, as was mentioned, this level is uncertain even in the Texas type sections, because most characteristic representatives of the Wordian assemblage (genus *Waagenoceras*) appear inside the Roadian Stage (in its current understanding). Second, the first occurrence of the assemblage with *Waagenoceras* species in the Tethyan sections is established inexactly. I can only mention that the undoubtedly Wordian ammonoids are unknown below the upper half of the Murgabian Stage.

As for conodonts and their potential for recognition of the Roadian Stage in the Tethyan region, I should note the following.

In the currently accepted version of the global scale, the lower and upper boundaries of the stage are placed at the first occurrence levels of *Jinogondolella nankingensis* and *J. aserrata*, respectively. However, the characteristic genera and species of the stage are not indicated. Inasmuch as the species mentioned above dwelt only in warm water environments, they lack in deposits accumulated in colder climatic zones, or in relatively deeper settings. Consequently, the possibility of recognizing the stage in question remains ambiguous without versatile characteristics of conodonts typical of the stage. In the full measure, this is evident from variable interpretations of the Roadian Stage boundaries in the Tethyan sections, which have been suggested by experts on the conodont biostratigraphy.

The index species of the *Jinogondolella nankingensis* conodont zone that presumably corresponds to the Roadian Stage is known in the Tethyan region only from few sections of South China, and thus, it is of a limited significance for the correlation purpose. To solve the problem, we should define the exact position of the above conodont zone relative to the fusulinid zonation easily applicable for correlation of the Tethyan sections. Such an attempt is exemplified by the subdivision scheme of the Luodian section in the Guizhou Province of South China (*Excursion Guidebook*, 1994; Zhu and Zhang, 1994; Jin, 1998a, 1998b; Henderson *et al.*, 1999). In this section (Fig. 2), *J. nankingensis* is noted first in Bed 46, i.e., 150 m above the first occurrence level of the index species (*Neoschwagerina simplex*, Bed 28) of the lower zone of the Murgabian Stage and only 25 m below the first occurrence of the genus *Yabeina* (Bed 49) characteristic of the Midian Stage. The overlying Bed 50 yields first speci-

mens of *J. aserrata* representing the index species of the Wordian Stage according to the new definition of the latter (*Excursion Guidebook*, 1994). In addition to conodonts, Bed 46 yields fusulinids *Sumatrina annae* Volz., *Afghanella megasphaerica* Leven, *Pseudodoliolina pseudolepida* (Deprat), and others species characteristic of the upper half of the Murgabian Stage.

Using these data and formally following the Roadian Stage definition on the basis of conodonts, Mei Shilong and Ch.M. Henderson (Mei *et al.*, 1999; Henderson *et al.*, 1999) argued that the Roadian Stage in the considered section should include Beds 46–49 that are 78 m thick in total. In the Tethyan scale, this interval corresponds to the uppermost part of the Murgabian Stage that yields the highly evolved *Neoschwagerina*, *Afghanella*, and *Sumatrina* species and to the lowermost Midian beds with *Yabeina* forms. In this case, the greater part of the Murgabian Stage and the entire Kubergandian Stage are referred to the Kungurian Stage of the Lower Permian, and the boundary between the Cisuralian and Guadalupian series is placed inside the Murgabian stage. Such a conclusion substantiated only by occurrence of a sole conodont species (*J. nankingensis*) recognized in some sections is absolutely inconsistent with the data on ammonoid distribution mentioned above. In addition, the boundary between Permian series, when it is placed inside the Murgabian Stage, seems doubtful from the standpoint of successive events, because nothing happened in the fusulinid community at this level. Accordingly, this boundary position is artificial and virtually unrecognizable at least in the Tethyan region.

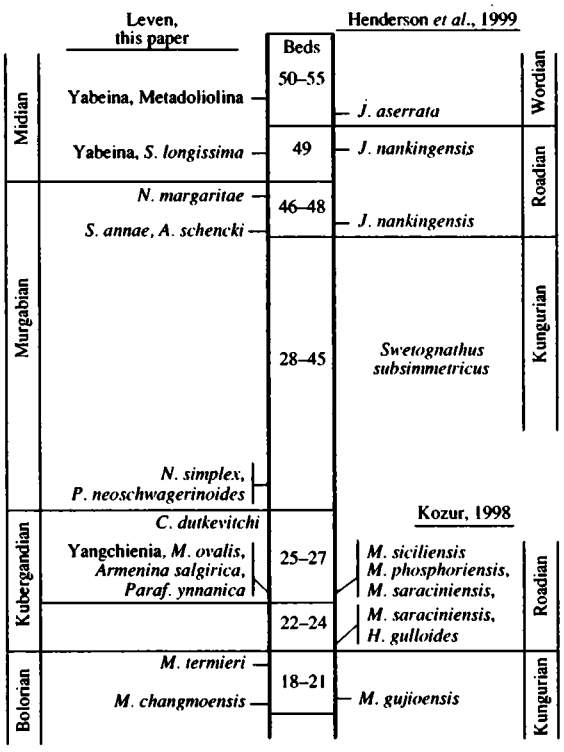

Fig. 2. Subdivision of the Luodian section in accordance with the Tethyan fusulinid (left) and standard conodont (right) scales. Fusulinids: (S.) *Sumatrina*; (N.) *Neoschwagerina*; (A.) Afghanella; (P.) *Presumatrina*; (C) *Cancellina*; (M.) *Misellina*; (Paraf.) *Parafusulina*. Conodonts: (M.) Mesogondolella; (H.) *Hindeodus*; (J.) *Jinogondolella*.

Kozur (1998) who had a possibility to get acquainted with conodont collection from the Luodian section criticized the conclusions of Mei and Henderson. In opinion of Kozur, the lower boundary of the Roadian Stage in that section should be placed at the appearance level of several smooth and more cryophilic (or deeper dwelt) gondolellids rather than at the first occurrence level of *J. nankingensis* that is probably redeposited. Like the latter, these gondolellids are descendants of *Mesogondolella idahoensis*. They appeared at the Kungurian–Roadian boundary and gave rise to two separate lineages: *M. idahoensis–M. saracinensis–M. siciliensis* and *M. idahoensis–M. phosphoriensis*. Using this concept, Kozur revised the age of beds with conodonts in the Luodian section. His stratigraphic scheme almost completely coincides with mine based on distribution of fusulinid taxa listed in publications (Zhu and Zhang, 1994; *Excursion Guidebook*, 1994) and identified by myself in samples collected during the excursion of the International Symposium on Permian Stratigraphy, Guiyang, 1994. Comparison of these fusulinid- and conodont-based schemes is shown in Figure 2. For instance, *Brevaxina* species and primitive *Misellina* forms, such as *B. dyhrenfurthi* (Dutkevich, *M. termieri* (Deprat), and *M. changmoensis* Dong, from beds 18 and 21b unambiguously indicate the Bolorian age of the beds, although the latter also yields first specimens of *Parafusulina yunnanica* widespread in the Kubergandian Stage. The undoubtedly Kubergandian fusulinids (genus *Yangchienia*) appear in Bed 25a. Beds 22–24 lack fusulinids, but they probably belong also to the Kubergandian Stage. In Bed 25b, fusulinids are represented by *Armenina salgirica* A. M.-Maklay, *Misellina megalocula* Wang et Sun, *M. confragaspira* Leven, *Yangchienia* sp., and *Parafusulina yunnanica* Sheng typical of the lower *Armenina–Misellina ovalis* Zone of this stage.

Judging from conodonts, namely from presence of *Mesogondolella gujioensis* (Igo), Kozur attributed Bed 18 to the Bolorian Stage. Beds 19–21 are barren of conodonts, but Bed 22 yields *Gullodus catalonoi* (Gullo et Kozur), *Mesogondolella saracinensis* Gullo et Kozur, and *Hindeodus gulloides* Kozur et Mostler. These species are characteristic of the Roadian Stage, but some of them could have been appearing earlier. Accordingly, Kozur attributes Beds 22–24 to the Roadian, although with some reservations. In his opinion, undoubtedly Roadian is Bed 25b yielding *M. saracinensis* Gullo et Kozur, *M. phosphoriensis* (Youngquist, Hawley et Miller), and *M. siciliensis* (Kozur) (in the upper part). As can be seen, the Kungurian–Roadian boundary in the Kozur's scheme almost exactly coincides with its position between the Bolorian and Kubergandian

stages of the Tethyan scale. As for the upper boundary of the Roadian Stage in the Loudian section, Kozur did not discuss the relevant data. In principle, he believes that the idea to consider the first occurrence of *J. aserrata* as marking the boundary in question is unfortunate, because this species populated only surface waters of warm sea basin and, thus, its correlation potential is very low (Kozur, 1998).

Summarizing data on conodonts from the Luodian section, I can state the following.

(1) The lower boundary of the Roadian Stage, when it is placed at the occurrence level of cryophilic (deep-water) smooth gondolellids, coincides with the lower boundary of the Kubergandian Stage of the Tethyan scale. As is shown above, ammonoid assemblages from the Kubergandian Stage stratotype lead to a similar inference.

(2) When this boundary in the Luodian section is defined by the first occurrence level of *J. nankingensis*, it falls into the upper part of the Murgabian Stage. In this case, the Kungurian Stage time-range exorbitantly increases so, as to span, in addition to the Bolorian Stage, the Kubergandian Stage and the greater part of the Murgabian Stage. Since such a correlation is unlikely, one should assume that either *J. nankingensis* is redeposited in the Luodian section, or it appeared in the Tethys later than in the Delaware basin of North America. In the last case, this species is unsuitable at all to represent a marker of the lower boundary of the Roadian Stage, in particular, and Guadalupian Series, as a whole.

Wordian Stage. Similarly to the Roadian Stage, this unit was established by Furnish (1973) on basis of ammonoids as corresponding to the *Waagenoceras* Genus-zone. Glenister *et al.* (1992) who proposed to include the Guadalupian Series with its stages into the global chronostratigraphic scale shared this standpoint. The lower stage boundary was suggested to coincide with the first occurrence of *Waagenocears* representatives inside the Pipeline Shales at the base of the Brushy Canyon Formation in the Guadalupe Mountains (Fig. 1). In the Glass Mountains, this level corresponds to the base of Willis Ranch Limestones in the middle part of the Word Formation, in traditional understanding of the latter. The boundary between the Wordian and overlying Capitanian stages was proposed to be at the first occurrence of the *Timorites* genus, the descendant of the *Waagenoceras* genus. In the section of the Guadalupian Mountains, first *Timorites* forms are recorded in the uppermost Manzanita Beds of the Cherry Canyon Formation (Furnish, 1973).

The current concept of the Wordian Stage range and boundaries distinguished on the basis of conodonts essentially differs from the original one. The stage is correlated with the *Jinogondolella aserrata* Zone having boundaries inside the Wordian and Capitanian stages of the former ammonoid-based scale. As was mentioned, the GSSP of the lower boundary of the Wordian Stage is placed slightly below the roof of the Getaway Limestones of the Cherry Canyon Formation in the section of the Guadalupian Mountains, where it coincides with the first occurrence of the index species. In the Glass Mountains, this boundary separates the Willis Ranch and Appel Ranch limestone units of the Word Formation (Glenister *et al.*, 1999).

Before discussing the problem of recognition of the Wordian Stage in the Tethyan region, I should note that this is as difficult as in the Roadian Stage case. The Wordian fusulinid assemblage is almost identical to the Roadian one. It is represented by endemics missing beyond the western Texas and, thus, unsuitable for distant correlations. Ammonoids and conodonts are rare in the corresponding interval of the Tethyan sections, and their finds are poorly correlated with fusulinid markers.

The typical Wordian (after Furnish) ammonoid assemblage is well represented in isolated limestone blocks of Sicily and Timor (Gemmellaro, 1887; Gerth, 1950). Some of its representatives are also recorded in the Far East, Tunisia, China, and Japan. The Marta assemblage described from the Kichkhi-Burnu locality of the Crimea (Tumanskaya, 1931) is also of the Wordian age. In the Tethyan region, deposits with Wordian ammonoids were traditionally correlated with the Murgabian Stage defined in the range of the *Neoschwagerina* Genus-zone. Indeed, when Wordian ammonoids associate with the fusulinid assemblage, the latter usually includes this index genus. At the same time, the *Neoschwagerina* stratigraphic range exceeds that of the Murgabian Stage, because representatives of this genus cross the Murgabian–Midian boundary. In sections reliably dated by fusulinids, *Waagenocears* representatives with sufficiently developed lobe lines coexist with advanced *Neoschwagerina* species that are absent below the upper Murgabian Stage. In Rupe de Passo di Burgio Sosio, the best known locality of Wordian ammonoids in Sicily, fusulinids (*Kahlerina, Rauserella, Dunbarula*, and *Sumatrina longissima* Deprat) characterize the basal part of the Midian Stage (Leven, 1996; Kozur and Davydov, 1996).

The Midian fusulinids are recorded together with Wordian ammonoids in Tunisia (*Dunbarula, Neoschwagerina, Yabeina*) and also in the Cache Creek locality of British Columbia, Canada (*Yabeina*) (Miller and Furnish, 1957; Ross, 1971). Thus, the reliably established stratigraphic range of the Wordian (after Furnish) ammonoid assemblage, first of all that of the *Waagenoceras* genus, comprises in the Tethyan region the upper half of the Murgabian Stage and the lowermost portion of the Midian Stage. Taking into consideration the opinion of Bogoslovskaya (1984) that the Marta ammonoid assemblage is similar to the Wordian one, it is not inconceivable that the stage range can be expanded to the base of the Murgabian. This assumption accords with data on diverse fusulinids from the Kichkhi-Burnu section, most of which characterize the lower zone of the Murgabian Stage, although some

Tethyan Scale (Leven, 1980)		Conodonts				Ammonoids	
		Henderson *et al.*, 1999		Kozur, 1998		Bogoslovskaya *et al.*, 1999	
Stages	Zones						
Dzhulfian (s. s.)	*Clarkina leveni*	Lopingian	Wuchapingian	Lopingian	Dzhulfian (s. l.)		Dzhulfian
	Araxoceras						
Midian	*Eoaraxoceras*						
	L. kumaensis *L. multiseptata, Y. globosa*	Guadalupian	Capitanian	Guadalupian	Capitanian	P_2	Capitanian
			Wordian				
	A. robbinsae, Y. archaica		Roadian		Wordian		Wordian
Murgabian	*A. schencki*						
	A. tereshkovae	Cisuralian	Kungurian		Roadian		Roadian
	P. neoschwagerinoides, N. simplex						
Kubergandian	*C. cutalensis*						
	Armenina, M. ovalis						
Bolorian	*M. parvicostata*				Kungurian	P_1	Kungurian

Fig. 3. Variants of correlation of the standard and Tethyan stages within the Guadalupian Series. Fusulinids: (M.) *Misellina*; (C.) *Cancellina*; (N.) *Neoschwagerina*; (P.) *Presumatrina*; (A.) *Armenina*; (Y.) *Yabeina*; (L.) *Lepidolina*.

samples yield also the middle and upper Murgabian species.

Conodonts *J. aserrata* and *J. postserrata*, which are now considered, respectively, as markers of the lower and upper boundaries of the Wordian Stage, are recorded beyond the United States only in few sections of South China. Therefore, their practical application for the Wordian Stage recognition in the Tethyan region still remains to be substantiated. Judging from fusulinids (*Yabeina, Metadoliolina*) occurring together with indicated conodont species in the Luodian section (Fig. 2), the Wordian Stage in its current conodont-based interpretation (Fig. 3) is fully located inside the Midian Stage of the Tethyan scale (*Excursion Guidebook*, 1994; Mei *et al.*, 1998). Nothing significant in fusulinid development is recorded near the stage boundaries in the conodont scale, and fusulinids are, therefore, useless for their recognition.

Inferences from above information are as follows.

(1) In the Texas type sections, Wordian Fusulinids are represented by endemic forms unsuitable for the stage recognition in Tethyan areas.

(2) The ammonoid assemblage of the Wordian Stage, in its new time-range shortened now, remains inexactly characterized, and this unit cannot be recognized, therefore, in the Tethyan sections (as well as in type sections of Texas) with the help of this fossil group. When the Wordian Stage is considered in its traditional range, some elements of its ammonoid assemblage can be found in particular Tethyan sections. Nevertheless, distribution ranges of respective genera and species are unclear. According to available data, they occur in the interval from the upper part of the Murgabian Stage to the lower part of the Midian Stage. It is likely as well that their lower distribution limit can be lowered down to the base of the Murgabian Stage.

(3) Judging from conodonts, the Wordian Stage corresponds to a part of the Midian Stage. Species marking new boundaries of the Wordian Stage are recorded only in the United States and South China being, thus, unsuitable for broad correlation with the Tethyan sections. The significance of conodonts decreases further, when one takes into consideration the fact that the guide assemblage of the stage is not established so far.

Capitanian Stage was established by Furnish (1973) as corresponding to the ammonoid *Timorites* Zone, though he did not indicate its stratotype and boundaries. Most representative cephalopod assemblages characteristic of this zone and, correspondingly, of the stage are recorded in Timor (the Basleo locality) and Mexico (the Las Delicias locality). In the Guadalupe Mountains, where the recently accepted type section of the stage is situated, the *Timorites* species occur in the Manzanita Beds of the Cherry Canyon Formation and in the Hegler Beds of the Bell Canyon Formation (Fig. 1). Despite the Hegler Beds Position at the very base of the Bell Canyon Formation, the latter is attributed almost entirely, except for the Lamar Limestones crowning it, to the Capitanian Stage. Furnish ascribed these limestones to the Amarassian Stage characterized by cephalopods known from the Timor and Mexico localities. The most peculiar feature of the Amarassian ammonoid assemblage is the first appearance of the genus *Cyclolobus* and xenodiscids. The Lamar Limestones proper are barren of ammonoids and their correlation with the Amarassian Stage is ambiguous to a considerable extent. It should be noted, however, that first xenodiscids (the genus *Xenaspis*) were recently discovered in the Pinery Beds of the Bell Canyon For-

mation, i.e., significantly below the Lamar Limestones (Wardlaw, 1996).

According to the accepted definition of the stage under consideration, its lower boundary (GSSP) is established at the first occurrence level of the *Jinogondolella postserrata* near the top of the Pinery Limestones, i.e., at the appearance level of first xenodiscids above beds yielding first *Timorites* specimens (Fig. 1). The GSSP of the upper stage boundary is defined in South China (Jin *et al.*, 1998a). In the Guadalupe Mountains, its level is close to the roof of the Lamar Limestones, where, like in South China, *Jinogondolella* species are replaced by the *Clarkina* forms (Kozur, 1998; Mei *et al.*, 1998).

Correlating the Capitanian Stage of the new global standard with units of the Tethyan scale, we again meet difficulties related to recognition of the stage boundaries (particularly, of the lower one) because of rare occurrence of conodonts and ammonoids, position of which in the fusulinid zonation remains unclear. Indeed, the Capitanian ammonoids (first of all, *Timorites* species) are recorded only in some Tethyan sections, where they are not accompanied by fusulinids. The best we can do, therefore, is to assume the Midian age of enclosing deposits. In addition, we should keep in mind that the stratigraphic range of *Timorites* genus is wider that the Capitanian Stage range inferred now on the basis of conodont biostratigraphy. As for conodont species *J. postserrata*, whose appearance is proposed to be marker of the lower stage boundary, they are recorded, in addition to sections in Texas, only in several sections of South China, where they occur in the upper half of the Midian Stage, as one can judge from accompanying fusulinids.

Thus, the available data indicate that the Capitanian Stage can be recognized only in some Tethyan sections of South China, where conodonts depict its lower and upper boundaries. In other cases, one can suggest only an approximate correlation of the stage with the uppermost part of the Midian Stage. Fossils, which could substitute conodonts by recognition of the lower stage boundary, are not established so far. As for the upper boundary of the Capitanian Stage, it is more or less coincident with the major biotic crisis resulted in extinction of larger fusulinids, almost all goniatites and corals, and of many other representatives of marine biota (Jin *et al.*, 1994; Leven and Korchagin, in press). Accordingly, this boundary is easily recognizable in the Tethyan region, although some discrepancies, those of the principal character included, exist as well. For instance, the GSSP of this boundary is officially placed at the first appearance level of *Clarkina postbitteri* in one of the South China sections (Jin *et al.*, 1998a, 1998b). Kozur (1998) suggests either to place it slightly below, at the base of the *Clarkina altudaensis* Zone, or, to the contrary, to draw it substantially higher, at the base of the *Clarkina leveni* Zone. In any case, any of the proposed levels can serve as the upper boundary of the Capitanian Stage and, consequently, the Guadalupian–Lopingian boundary is also well recognizable in the Tethyan region.

The Lopingian Series with its Wuchapingian and Changhsingian stages is included into the global geochronological scale according to the proposal of Chinese stratigraphers; all these units are established in the South Chinese (i.e., Tethyan) sections. Their boundaries are well marked by changes in the conodont, fusulinid, and ammonoid assemblages and, thus, the recognition of the units meets no serious problems in the entire Tethyan realm. Tracing them beyond the Tethyan region, one meets difficulties, but this problem is out of the scope of this work.

CONCLUSION

The estimated correlation potential of the standard chronostratigraphic scale in the Tethyan region leads to the following conclusions.

(1) In principle, it seems reasonable to subdivide the Permian System into three series, because the three-member scale reflects adequately the natural–historical development of the biosphere. At the same time, criteria used to define the lower boundary of the Guadalupian Series need to be specified or revised. The available data show that in case, when this boundary is correlated as proposed, with the first occurrence of thermophilic (shallow-water) *J. nankingensis*, it falls inside the Murgabian Stage of the Tethyan scale, at the level near the stage top, where nothing happened with marine biota. Situation is quite different, when the lower boundary of the Guadalupian Series is placed at the first occurrence level of cold-resistant conodonts and ammonoids. In this case, it coincides with the base of the Kubergandian Stage of the Tethyan scale marked by essential renewals in marine fossil assemblages. Accordingly, the lastly mentioned level can be considered as better corresponding to the boundary between the Cisuralian and Guadalupian series. The noted discrepancies are probably related to the redeposited state of *J. nankingensis* in the South Chinese sections. Additional search for this species in lower layers of these sections are desirable therefore.

(2) Asselian, Sakmarian, Artinskian, and Kungurian stages of the lower Cisuralian Series of the standard scale correspond to the Asselian, Sakmarian, Yakhtashian, and Bolorian stages of the Tethyan scale, respectively. Nevertheless, the inadequate characteristics of conodonts and ammonoids from the Tethyan sections, as well as present-day disagreements concerning the definition criteria of stage boundaries in the standard scale, bring the above statement into challenge. Despite this situation, it is quite possible to practice the indicated stages of the standard scale instead of the Tethyan stages. The sole exception is the Bolorian Stage, since its correspondence to the Kungurian Stage remains uncertain, if we take into consideration the aforemen-

tioned controversial positioning of the lower boundary of the Guadalupian Series. The problem solution needs in additional data that would confirm the boundary position in Tethyan sections at the base of the Kubergandian Stage, but not in the uppermost part of the Murgabian Stage.

(3) None of the stages of the middle Guadalupian Series is recognizable in the Tethyan region. At the first approximation, it can be assumed that the Roadian, Wordian, and Capitanian stages of the standard scale correspond respectively to the Kubergandian, Murgabian, and Midian stages in the Tethyan region, if their correlation would be based on ammonoids according to tradition. Correlation with the help of ammonoids is impossible at present because of their rare occurrence. The current stage boundaries in the standard scale are inferred from the conodont biostratigraphy and substantially differ from those corresponding to biotic events in ammonoid communities. According to the available data, all boundaries between three stages of the standard Guadalupian Series fall in the Midian–uppermost Murgabian interval of the Tethyan scale, when one strictly follows the conodont-based boundary definitions. It is apparent, therefore, that the Roadian, Wordian, and Capitanian stages are absolutely unsuitable for the Tethyan stratigraphy, because their boundaries in current interpretation have no confirmation in fusulinid zonation that is critical for recognition of the Tethyan stages. In addition, the Kungurian Stage time-range, which should correspond in this case to the Bolorian and Kubergandian stages coupled with the greater part of the Murgabian Stage, will be exorbitantly large.

(4) Two stages of the upper Lopingian Series of the Permian System are established in South China. Naturally, there are no problems with their application in the Tethyan region, particularly if we take into consideration that their paleontological characteristics are known much better than those of the Guadalupian stages. Some discrepancies concerning the lower boundary of the series are insignificant and can be eliminated already now by voting. The Wuchapingian Stage time-range in the standard scale would exceed that of the Dzhulfian Stage in the Tethyan scale, if its lower boundary is correlated with the first occurrence level of *Araxoceras* and *Clarkina leveni*. The Changhsingian Stage of the standard scale exactly coincides with the Dorashamian Stage of the Tethyan scale.

(5) The aforesaid indicates that the standard stages of the Cisuralian and Lopingian series of the Permian System can be applied in the Tethyan stratigraphy after some coordination and specifications. Stages of the Guadalupian Series are still inappropriate despite the fact that they are approved by the International Subcommission on Permian Stratigraphy. This is a result of new approach, when, instead of substantiation of the Guadalupian stages on the basis of versatile paleontological characteristics and changes in faunal assemblages at their boundaries, the procedure was reduced to formal tracing of the species succession in the sole phylogenetic lineage of conodonts. The formalized recognition of stages and their boundaries on the basis of the single criterion (appearance of successive conodont species) reduces the possibility of distinguishing the same units in other regions.

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Biostratigraphy of the Marine Paleogene in the West Siberian Plate

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Abstract—The regional stratigraphic scheme suggested for the marine Paleogene deposits in the West Siberian plate is based on investigation results of macro- and microfossils (foraminifers, dinoflagellates, radiolarians, diatomaceous algae, silicoflagellates, Elasmobranchii, ostracodes, pollen, and spores). In the West Siberian plate, like in many structural–facies zones of the East European platform, there was a hiatus in sedimentation at the Cretaceous–Paleogene boundary. The Talitsa Horizon corresponds here to the Danian and Selandian presumably coupled with the initial Thanetian. The lower Lyulin-Vor Subhorizon is dated back to the late Paleocene; the middle Lyulin-Vor Subhorizon is correlated with the Ypresian, and the upper one is attributed to the interval spanning the second half of the Ypresian and Lutetian. The Tavda Horizon of the Bartonian–Priabonian age presumably comprises as well the terminal Lutetian. The Kurgan Beds (s. str.) representing marine facies of the Atlym Horizon correspond in the type area (near the town of Kurgan) to the lower Oligocene. Boundaries of these lithostratigraphic units are diachronous, since the Paleoarctis in the north and the Tethys in the south were connected with the epicontinental West Siberian sea basin producing an irregular impact on sedimentation in the latter.

Key words: Paleocene, Eocene, Oligocene, West Siberian plate; foraminifers, dinocysts, radiolarians, diatoms, Elasmobranchii, mollusks, silicoflagellates, ostracodes, spore–pollen assemblages; Talitsa, Lyulin-Vor and Tavda horizons, Serov and Irbit formations.

INTRODUCTION

This work was stimulated by the meeting of the Regional Interdepartmental Stratigraphic Commission on the Paleogene System (May 29–31, 2000), where the problems of the marine Paleogene biostratigraphy in the West Siberian plate were discussed in connection with the Regional Stratigraphic Conference scheduled for the fall of 2000. The commission considered two projects of stratigraphic schemes prepared under supervision of V.A. Martynov and V.M. Podobina. Age ranges of the same litho- and biostratigraphic units of the marine Paleogene were differently interpreted in these schemes, and proper critical remarks led the commission members to the conclusion that one more regional version of the scheme is desirable in order to take into account the most recent biostratigraphic data. The present group of authors includes experts from Moscow, St. Petersburg, and Yekaterinburg. They all have worked for some or other period and still are working on the Paleogene biostratigraphy in West Siberia, trans-Urals, Pechora basin, and Turgai depression, being involved as well into elaboration and revision of regional zonations for different groups of fauna and flora. The regional stratigraphic scheme presented in this paper includes our common conclusions coordinated to a considerable extent.

HISTORICAL REMARKS

A.P. Karpinsky was first to discover in the 1880s the Paleogene marine deposits in the eastern trans-Urals and further eastward in the plate. He distinguished principal subdivisions of the section: a sequence of the Lower Paleogene clayey-siliceous rocks, another sequence of clayey-silty deposits, and brown sandstone and sand beds bearing remains of the Oligocene mol-

† Deceased.

lusks (Kurgan Beds). The comprehensive investigation of these deposits was initiated in the terminal 1940s in connection with the first stage of the regional geologial mapping and petroleum prospecting. The team of geologists and stratigraphers who participated in this work created the first (1956) and second (1960) more precise stratigraphic schemes of the marine Paleogene in West Siberia. Among those whose contribution to the recognition, paleontological substantiation, and dating of regional lithostratigraphic subdivisions was especially significant, we should mention here Z.T. Aleskerova, B.E. Antypko, A.K. Bogdanovich, S.A. Chirva, N.G. Chochia, L.D. Dain, A.E. Eremeeva, F.G. Gurari, A.P. Jouse, A.V. Khabakov, A.I. Krotov, V.V. Lavrov, P.F. Li, V.A. Lider, R.Ch. Lipman, M.I. Mandel'shtam, V.A. Martynov, V.P. Nikitin, V.A. Nikolaev, G.N. Papulov, I.M. Pokrovskaya, S.D. Rabinovich, N.N. Rostovtsev, N.V. Rubina, S.B. Shatskii, K.G. Shibkova, A.P. Sigov, N.N. Subbotina, M.V. Ushakova, T.F. Vozzhennikova, and I.G. Zal'tsman. When the first stage of regional investigations was accomplished in the earliest 1970s, its results were summarized by Martynov, Sigov, and Chirva who wrote a chapter on geology of West Siberia in the volume "Startigraphy of the USSR" (*Stratigrafiya...*, 1975). Additional generalizations were published in a series of periodicals, monographs, and other editions.

The first stage of investigations resulted in the proper definition of stratigraphic ranges of main lithostratigraphic subdivisions of the Paleogene, distribution areas of which were also outlined. As drilling operations were concentrated at that time in western areas of the plate (Gurari, 1996), these subdivisions and their type sections were recognized mostly in the trans-Urals. Aleskerova and Osyko (see in *Tezisy...*, 1956, p. 24) reported on the most complete section of the Talitsa Formation that was recovered near the eponymous settlement of the Sverdlovsk district. Sigov described stratotypes of the Serov and Irbit formations distinguished in the vicinity of eponymous towns (*Tezisy...*, p. 25). The Lyulin-Vor Formation widespread everywhere in the plate was originally characterized by Li (*Tezisy...*, p. 26), who named it after the eponymous hills located in the eastern slope of the Urals within the Sos'va River basin. Finally, the type sections of the Tavda Formation and Kurgan Beds are also situated in the trans-Urals near the eponymous towns (Bogdanovich, 1944; Rostovtsev, 1955; Panova, 1970).

The regional Paleogene biostratigraphy achieved a noticeable success in two decades after the war. It was a time when successive microplankton assemblages were distinguished in the West Siberian sections, e. g., those of foraminifers and radiolarians (works by Subbotina, Ushakova, Eremeeva, Kisel'man, Kipriyanova, Lipman and Gorbovets), or of diatoms, silicoflagellates and dinoflagellates (works of Jouse, Krutov, Rubina, Shibkova, and Vozzhennikova). Moreover, it was especially important that Subbotina and Ushakova discovered in West Siberia an assemblage of planktonic foraminifers comparable in taxonomic composition with these fossils from the *Lyrolepis* Beds of the northern Caucasus (at present, the Kuma Horizon of the terminal middle Eocene).

Beginning from the second half of the 1960s, biostratigraphic investigations of the West Siberian Paleogene were concentrated on the following topics: detailing of stratigraphic schemes that existed before; dating of lithostratigraphic units; taxonomic composition of various biotic groups and their zonal assemblages; correlation of the latter with their counterparts in neighboring and remote regions. Among scientists who contributed much to the Paleogene biostratigraphy of the trans-Urals and West Siberia in the last period, we should mention E.O. Amon, O.N. Vasil'eva, Z.I. Glezer, G.E. Kozlova, I.A. Kul'kova, I.A. Nikolaeva, V.M. Podobina, and N.I. Strel'nikova.

REGIONAL MATERIALS USED IN THE STRATIGRAPHIC SCHEME

The suggested version of stratigraphic scheme (Fig. 1) is elaborated by experts on the following fossil groups: foraminifers (Beniamovskii and Bugrova who used data of Subbotina, Kisel'man, Ushakova and Podobina in addition to their own materials), diatomaceous algae (Glezer, Oreshkina, Radionova and Strel'nikova), radiolarians (Amon, Vitukhin and Kozlova), dinocysts (Aleksandrova, Vasil'eva, Zaporozhets and Yakovleva who used data of Kul'kova and Bakieva), and ostracodes (Nikolaeva). Also included in the scheme are paleocarpological zones after Nikitin (1998), palynozones after Panova (see in *Praktícheskaya...*, 1990), and zonation of Elasmobranchii after Zhelezko and Kozlov (1999). Data on malacofauna are presented in interpretation of O.V. Amitrov, I.A. Goncharova and S.V. Popov from the Paleontological Institute, Russian Academy of Sciences. Beginning from the early 1990s, Akhmet'ev published a series of papers, where he discussed, in addition to other problems, possible climatic events in the northern hemisphere, when the West Siberian epicontinental paleobasin and Turgai strait were connected with the World Ocean.

In many cases, assemblages of various biotic groups are directly correlated on the basis of their parallel investigations in the same sections. For instance, the Eocene foraminifers, dinocysts, diatoms, spores, and pollen were simultaneously studied in the Kamyshlov Quarry section (Aleksandrova, Amon, Beniamovskii, Bugrova, Zaporozhets, Kozlova, Oreshkina and Radionova). The section of Borehole "K" in the Komsomol'skaya area to the west of the Khanty-Mansiisk was sampled at once for the analysis of radiolarians, diatoms, dinocysts, spores, and pollen (Glezer, Kul'kova, Panova, and Kozlova). The same groups of organisms were studied in the borehole sections 11, 29, and 32 of the Pur–Taz interfluve area (Glezer, Kul'kova, Panova and Yakovleva), and also in borehole sections 29 and 30 of the Sos'va River basin (Kozlova, Kul'kova,

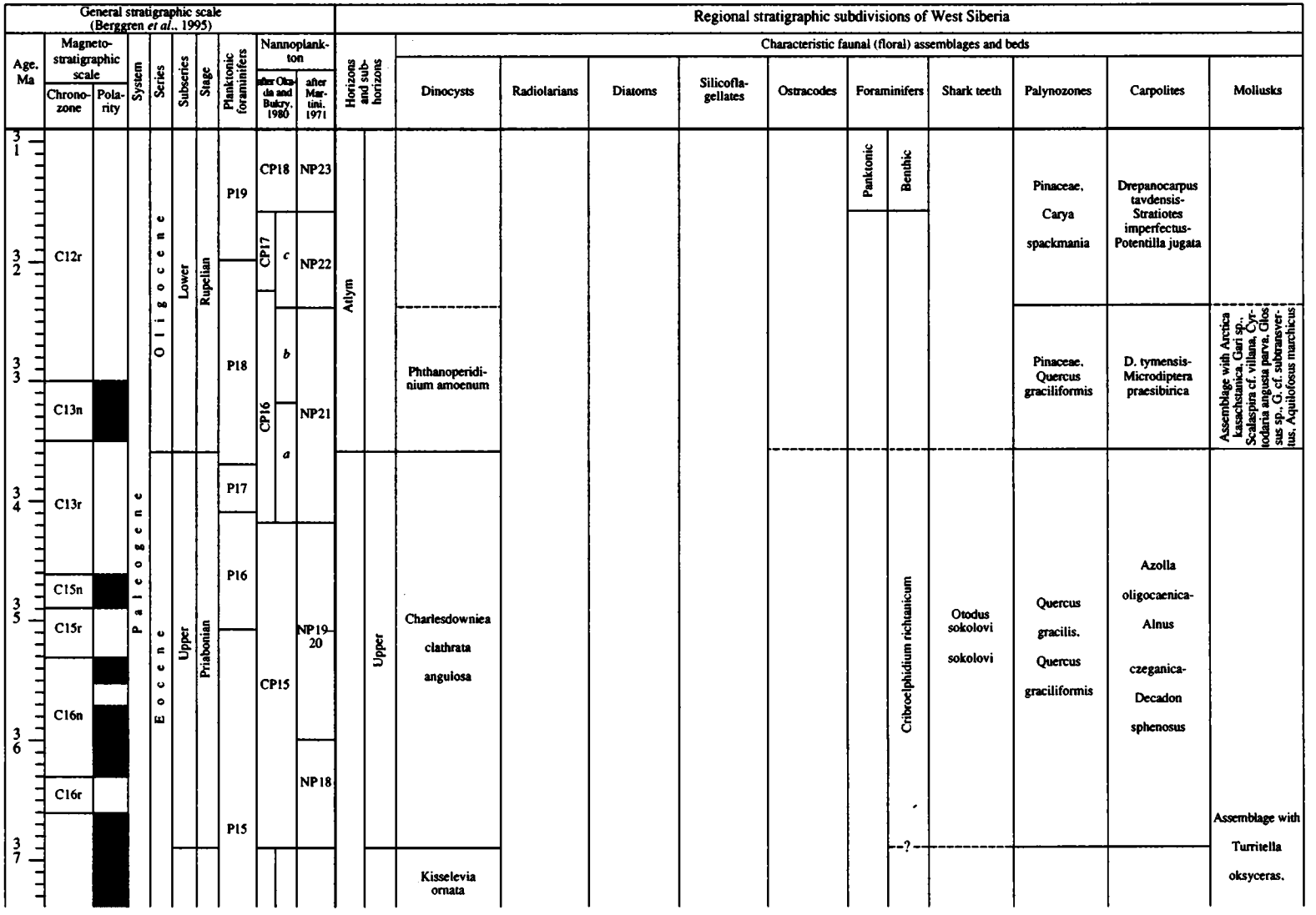

Fig. 1. Regional stratigraphic scheme of Paleogene deposits in West Siberia (dinocyst zonation after Aleksandrova, Vasil'eva, Zaporozhets, and Yakovleva with additions of Kul'kova, 2000; radiolarian zonation after Kozlova, 2000; diatom zonation in the left part of the column after Strel'nikova, Oreshkina, and Radionova, 2000, and in the right part after Glezer, 2000; silicoflagellate zonation after Glezer, 2000; foraminiferal zones and beds after Beniamovskii with additional data from Ushakova, 1959; *Foraminifery...*, 1964; Kisel'man, 1978; Podobina, 1998; ostracode zonation after Nikolaeva, 2000; Elasmobranchii zonation after Zhelezko, 1999; palynozones after Panova, 1990; and carpological zonation after Nikitin, 1998).

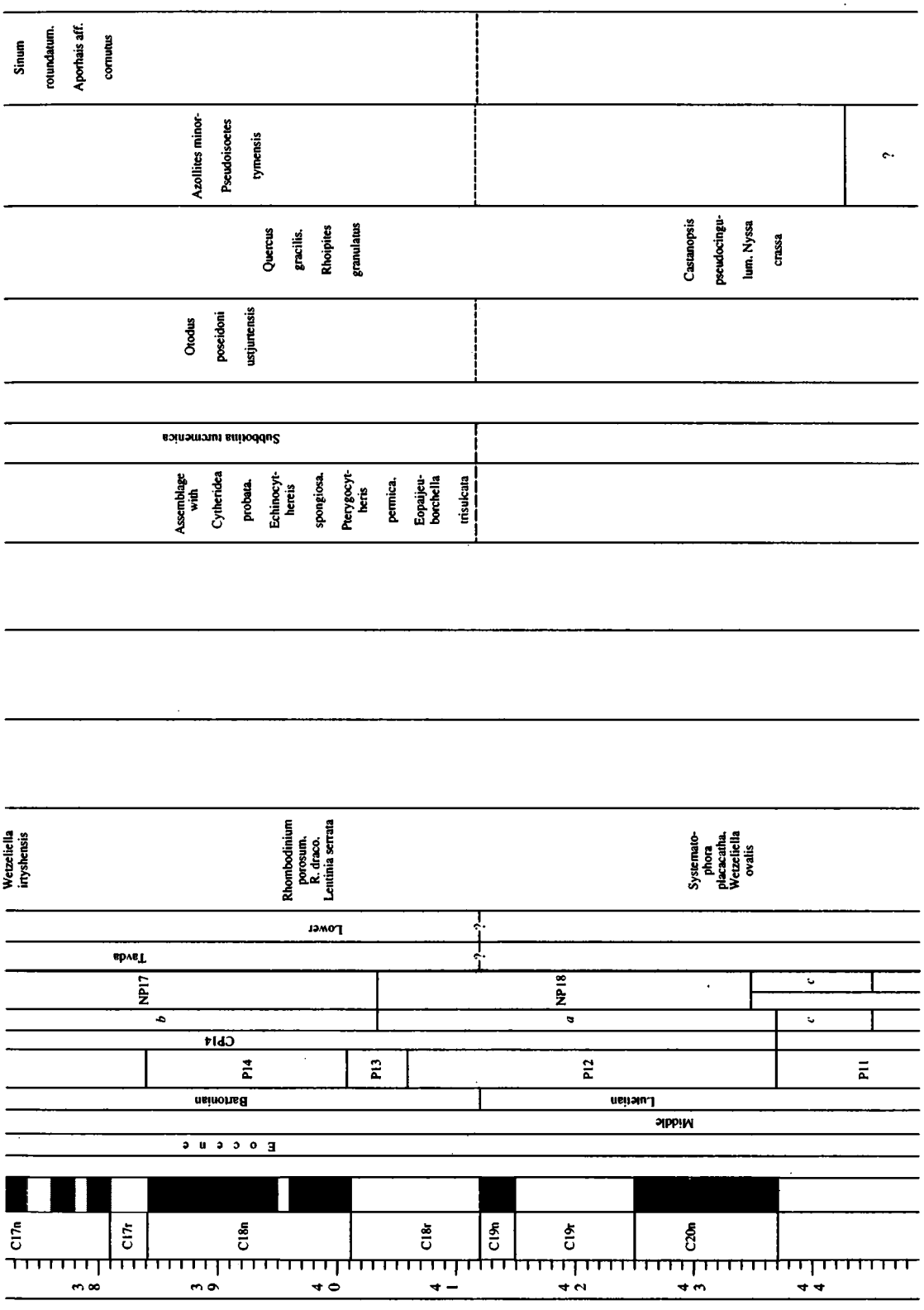
Wetzeliella intyshensis
Rhombodinium porosum. R. draco. Lentinia serrata
Systematophora placacatha. Wetzeliella ovalis
Lower
Tavda
NP17
NP18
b
a
c
CP14
P14
P13
P12
P11
Bartonian
Lutetian
Middle
Eocene
C17n
C17r
C18n
C18r
C19n
C19r
C20n
38
39
40
41
42
43
44
Assemblage with Cytheridea probata. Echinocytheris spongiosa. Pterygocytheris pernica. Eopaijenborchella trisulcata
Subbotina lozmenica
Otodus posidoni usjuntensis
Quercus gracilis. Rhoipites granulatus
Azollites minor-Pseudoisoetes tymensis
Sinum rotundatum. Aporrhais aff. cornutus
Castanopsis pseudocinguilum. Nyssa crassa
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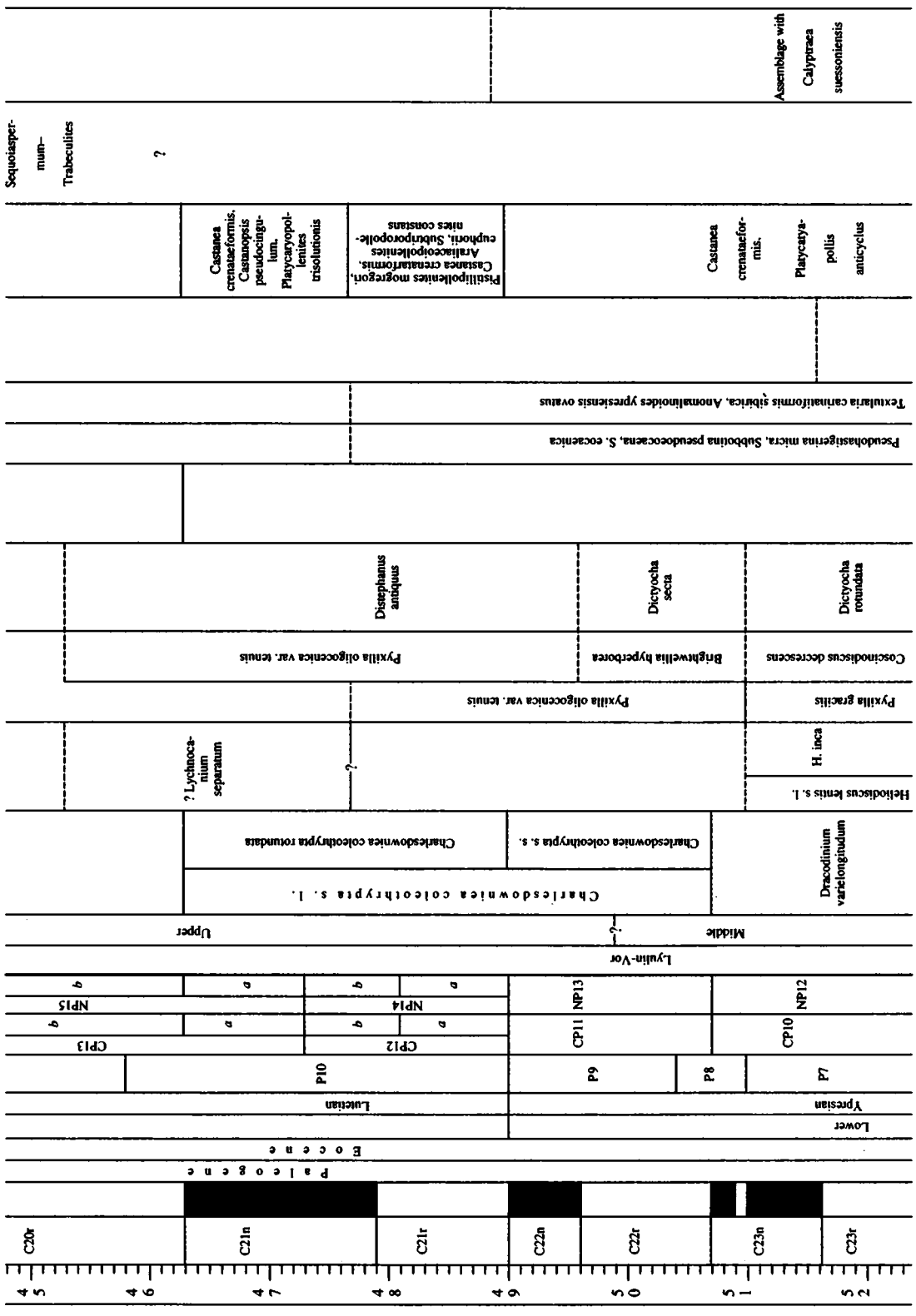
Fig. 1. (Contd.)


Assemblage with Calyptraea suessoniensis
Sequoiaspermum–Trabeculites
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Pistillipollenites mcgregorii, Casuarinidites crenaiformis, Arecipites echinatus, Arecipites punctatus, Arecipites sp., Arecipites sp. 1, Arecipites sp. 2, Arecipites sp. 3, Arecipites sp. 4, Arecipites sp. 5, Arecipites sp. 6, Arecipites sp. 7, Arecipites sp. 8, Arecipites sp. 9, Arecipites sp. 10, Arecipites sp. 11, Arecipites sp. 12, Arecipites sp. 13, Arecipites sp. 14, Arecipites sp. 15, Arecipites sp. 16, Arecipites sp. 17, Arecipites sp. 18, Arecipites sp. 19, Arecipites sp. 20, Arecipites sp. 21, Arecipites sp. 22, Arecipites sp. 23, Arecipites sp. 24, Arecipites sp. 25, Arecipites sp. 26, Arecipites sp. 27, Arecipites sp. 28, Arecipites sp. 29, Arecipites sp. 30, Arecipites sp. 31, Arecipites sp. 32, Arecipites sp. 33, Arecipites sp. 34, Arecipites sp. 35, Arecipites sp. 36, Arecipites sp. 37, Arecipites sp. 38, Arecipites sp. 39, Arecipites sp. 40, Arecipites sp. 41, Arecipites sp. 42, Arecipites sp. 43, Arecipites sp. 44, Arecipites sp. 45, Arecipites sp. 46, Arecipites sp. 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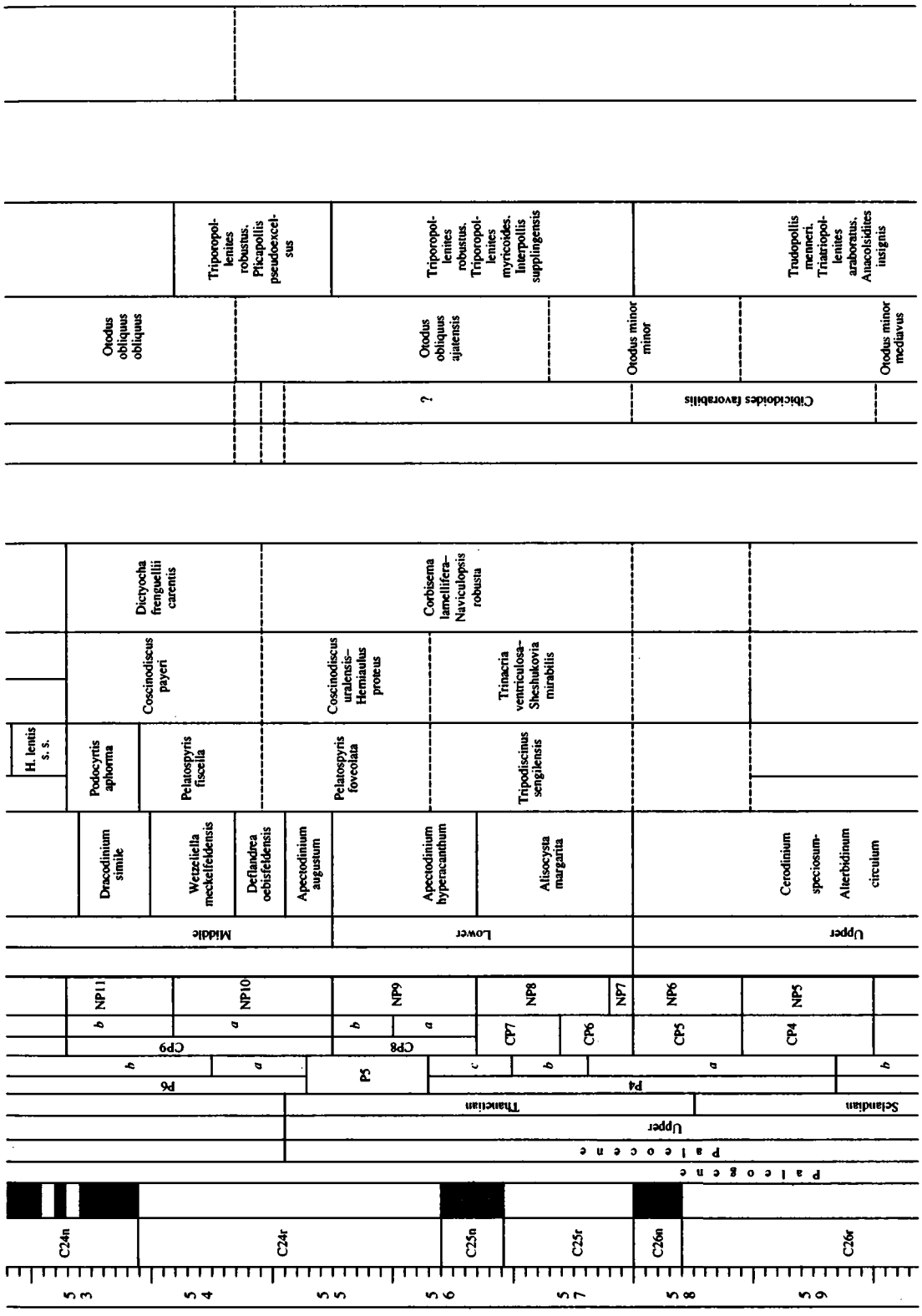
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Tudopollis menneri, Triatriopollenites araboratus, Anacolsidites insignis
Otodus minor mediavus
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Cibicidoides favorabilis
Dictyocha frenguellii carentis
Corbisema lamellifera–Naviculopsis robusta
Coscinodiscus payeri
Coscinodiscus uralensis–Hemiaulus proteus
Trinacria ventriculosa–Sheshukovia mirabilis
H. lentis s. s.
Podocyritis aphorma
Pelatospyris fiscella
Pelatospyris foveolata
Tripodiscinus sengilensis
Dracodinium simile
Wetzeliella meckelfeldensis
Deflandrea oebisfeldensis
Apectodinium augustum
Apectodinium hyperacanthum
Alisocysta margarita
Cerodinium speciosum–Alterbidinium circulum
Middle
Lower
Upper
NP11
NP10
NP9
NP8
NP7
NP6
NP5
b
a
CP9
CP8
CP7
CP6
CP5
CP4
P6
P5
P4
Thanetian
Selandian
Upper
Paleocene
C24n
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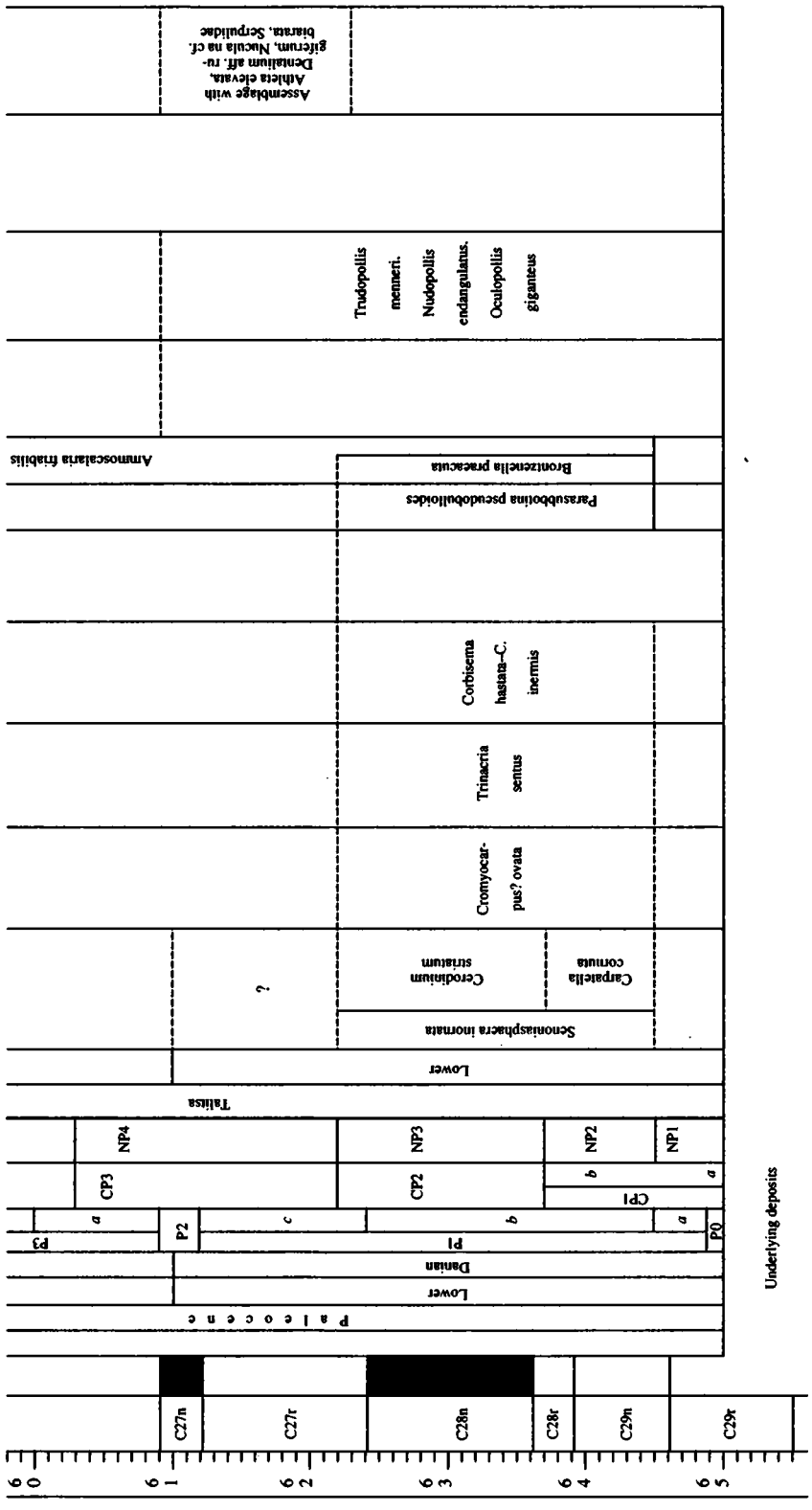
Fig. 1. (Contd.)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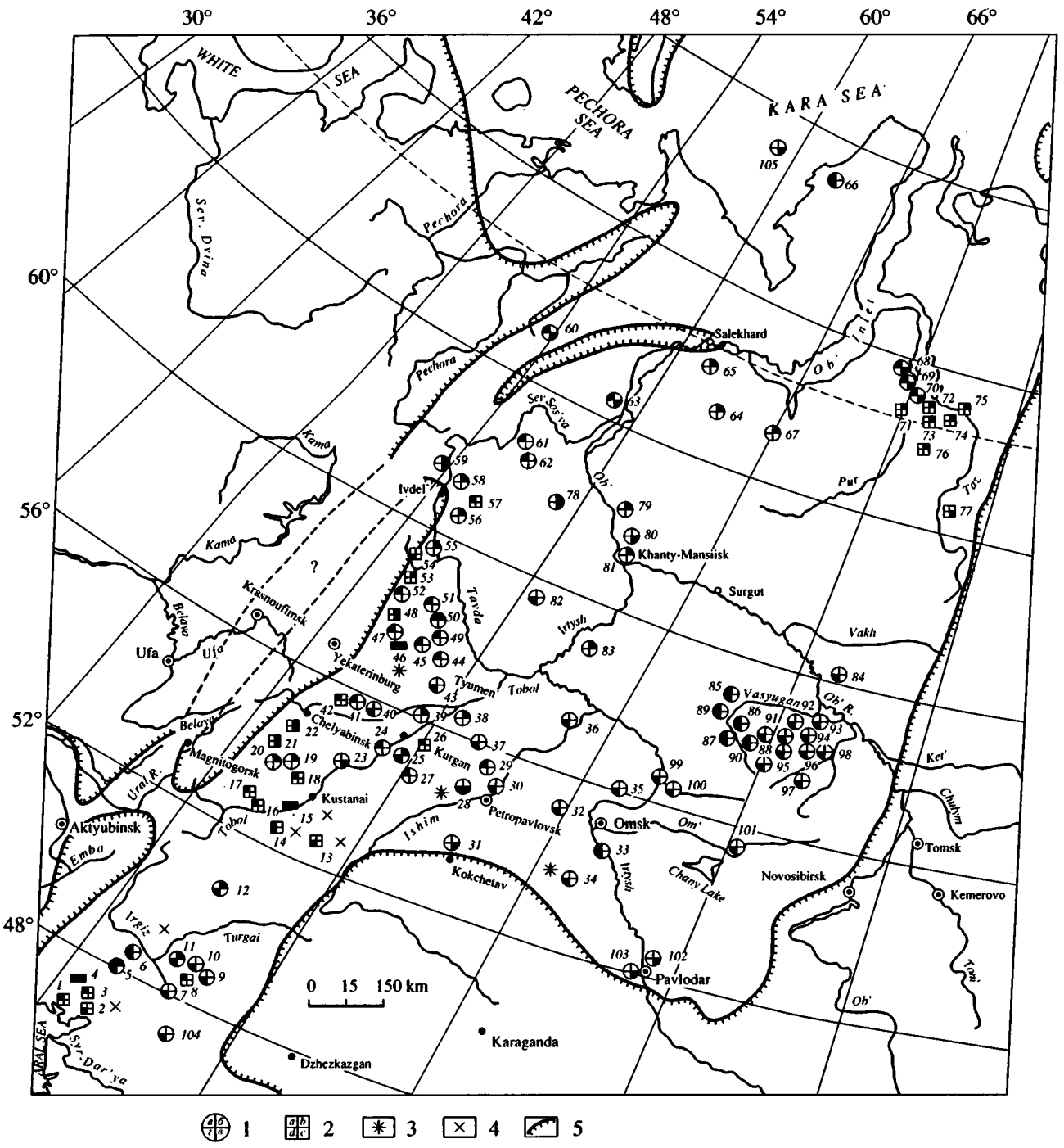
NP2
NP1
Talisa
Lower
Schonaisphaera inornata
Carpatella cornuta
Cerodinium striatum
?
Cromyocarpus? ovata
Trinacria sentus
Corbisema hastata–C. inermis
Parasubbotina pseudobulloides
Bronzenella praecursa
Ammoscalaria friabilis
Trudopollis menneri.
Nudopollis endangulatus.
Oculopollis giganteus
Assemblage with Ailanthes elevata, Denaeum aff. rugiferum, Nuculana cf. biarata, Serpulidae
Underlying deposits

Fig. 1. (Contd.)



Strel'nikova and Yakovleva). Dozens of other sites were investigated in the same manner, and most important sections studied by authors are shown in the map (Fig. 2) supplementing the suggested scheme. In our work, we accept the structural-facies zoning suggested by Shatskii (1978) for the Paleogene deposits in the West Siberian plate (Fig. 3).

Substantiating our conclusions, we used most recent data on microplankton groups from Paleogene deposits of the East European platform, Crimea–Caucasus region, Turan plate, and Central Asia. Data on inner and marginal parts of the North Sea basin of Western Europe are also used in our synthesis, as all stratotypes of Paleogene stages, except for that of the Priabonian, are concentrated there. Materials obtained from the West European sections turned out to be useful not only for correlation between the West Siberian Paleogene subdivisions and stages of the standard scale, but also for the comparative taxonomic analysis of microplankton groups from both regions. The concept of the Inter-

national Stratigraphic Commission to use the Global Boundary Stratotype Section and Point was principal in our evaluation of boundary ages (limitotypes) of the standard Paleogene subdivisions. In this work, we accepted three of these datum levels: 65 Ma for the Cretaceous–Paleogene boundary, 33.7 Ma for the Eocene–Oligocene boundary, and 23.8 Ma for the Paleogene–Neogene boundary. Investigation results of West European stratigraphers who studied Tertiary basins in northwestern Europe (IGCP project no. 124), as well as data of Russian Working Groups on the Paleogene problems (IGCP projects nos. 174, 308 and 326), were used in the full measure in our stratigraphic scheme. In addition, some of the present authors participated the Goteborg Conference of 1999, where the Paleocene–Eocene boundary interval was considered in various aspects. After the conference, there was a chance to visit and sample the Danian and Selandian stratotypes, and the sampled microplankton groups were subsequently compared with the corresponding West Siberian microfossils. Finally, we used all achievements of the sequence stratigraphy and "chronobiostratigraphy," which concern the Paleogene deposits of petroliferous basins and inner depressions of northwestern Europe (*Chronostratigraphy...*, 1998). As is known, the team of paleontologists from English, Dutch, French, and Norwegian petroleum companies outlined the key significance of dinocysts for the detailed subdivision of the epicontinental Paleogene. Owing to this fact, they effectively correlated the corresponding zonal scales elaborated using siliceous, calcareous, and organic-walled microfossils.

Turning back to the Paleogene biostratigraphy in West Siberia, we should emphasize once again that it was rationally developing beginning from the post-war time, when stratigraphers tried to correlate the distinguished regional stratigraphic units (horizons and formations) with stages and substages of the standard scale using zonal assemblages of benthic foraminifers, diatoms, radiolarians, silicoflagellates, spores, and pollen, although such an approach met certain difficulties in the 1950–1960s. Actually, the first zonal scales for siliceous plankton from open sea basins were suggested only in the 1970s. The diatom and radiolarian zonations established in West Siberia were most detailed at that time and served not infrequently as standards. Moreover, investigations of siliceous and calcareous microfossils developed separately in that period. Only in the 1980–1990s, when Strel'nikova, Kozlova and Glezer revised successions of siliceous microfossils and Kul'kova investigated distribution of dinocysts, the stratigraphic range of siliceous deposits in West Siberia that initially corresponded to the middle–upper Eocene was extended to the Paleocene–lower Lutetian (?) interval. Unfortunately, provincialism of benthic foraminifers from Siberia was always an obstacle for using them in stratigraphic correlations. Age constraints based on this fossil group were inferable just via their comparison with foraminiferal assemblages from Kazakhstan and Central Asia. Only these assemblages allowed linking of West Siberian foraminiferal groups with those characterizing the regional stages of the Crimea–Caucasus region and the standard stages of the International scale.

The outlined and some other reasons are responsible for the fact that age ranges of many Paleogene sequences, as they are defined now and in "resolutions" of the Stratigraphic Conference of 1976, remain to be at variance. Correlation between the general and national stratigraphic scales became possible only after the resolution of the Subcommission on the Paleogene System of the International Stratigraphic Commission, when a scheme of the Paleogene stage subdivisions was suggested. In the late 1980s, the resolution was authorized by the 28th session of the International Geological Congress. The middle Eocene status was attributed to the Bartonian Stage, and the Paleocene was subdivided into two subseries: the lower corresponding to the Danian Stage and the upper one comprising the Selandian and Thanetian stages. In the former stratigraphic scale of the Paleogene that was accepted in the USSR, the regional Bodrak Stage of the Crimean scale was considered as an equivalent of the middle Eocene. This regional stage was correlated with the Bartonian

Fig. 2. Geographic scheme of Paleogene sections with studied microfossils: (1) core sections and (2) exposures, which yielded dinocysts, spores and pollen (a), radiolarian (b), diatoms and silicoflagellates (c), and foraminifers (d); (3) core sections with planktonic foraminifers *Subbotina turcmenica*; (4) core sections with nannoplankton; (5) distribution limits of marine deposits (Lithological-Paleogeographic Maps of Paleogene Deposits in the USSR, 1968, with modifications). Number in the scheme denote the following sites and boreholes (BH): *1*, Perovskii Bay; *2*, Terbenbes; *3*, Sary-Oba; *4*, Tas-Aran; *5*, BH 50; *6*, BH 52; *7*, BH 125; *8*, Chelkar-Nura; *9*, BH 96; *10*, BH 95; *11*, BH 85; *12*, BH 255; *13*, Kushmurun; *14*, Belinskii; *15*, Sokolovskii; *16*, Ayat II; *17*, Ayat III; *18*, Kachar; *19*, Semiozernoe, BH 7-k; *20*, Duvankul', BH 19-k; *21*, Kuchigino; *22*, Korkino; *23*, BH 72; *24*, Kurgan, Chelnokovo quarry; *25*, BH 1-ok; *26*, ZHBI-1,2; *27*, BH 36; *28*, Petukhovo; *29*, BH BK-49; *30*, Petropavlovsk; *31*, BH 15972; *32*, Oktyabr'skaya, BH 1-k; *33*, Achair, BH 011-BP; *34*, Ozernyi; *35*, Sargatskoe, BH 1-r; *36*, Vikulovo; *37*, Zavodoukovsk, BH 3-k; *38*, Novozaimka-Ishim, BH 7-k; *39*, Ingalinsk; BH 1-k; *40*, Shadrinsk; *41*, BH IG-32; *42*, Dolmatovo-Pershino; *43*, Tyumen', BH 1-r; *44*, Luchinkino; *45*, Romanovka, BH 8005; *46*, Kamyshlov; *47*, BH 7-d; *48*, Irbit; *49*, Elan', BH 8; *50*, Lipatovskoe, BH LD-1; *51*, Turinsk, BH 1-r; *52*, Vogulka, BH 91; *53*, Lyalya-Titovo (exp. 13); *54*, Shumikhinskoe; *55*, Serov; *56*, Lyavdinka, BH 28; *57*, Loz'vinskaya Pristan'; *58*, Severnaya Sos'va; *59*, Ust'-Man'ya, BH 19-u; *60*, BH 228; *61*, BH 29; *62*, BH 30; *63*, Berezovo, BH 1-r; *64*, Polui, BH 27; *65*, Numgi-Sambur, BH 47; *66*, Kharasavei; *67*, Nadym, BH 9-k; *68*, BH 32; *69*, BH 29; *70*, BH 11; *71–76*, exposures in the Pur–Taz interfluve; *77*, Pechal'ka; *78*, Komsomol'skaya, BH 14-k; *79*, Kazym, BH 4-k; *80*, Atlym; *81*, Kondinskoe, BH 2k; *82*, Leushi; *83*, Uvat, BH 1r; *84*, Paiduchinskaya, BH 1; *85*, BH 1; *86*, BH 2; *87*, BH 3; *88*, BH 5k; *89*, BH 1-R; *90*, BH 4; *91*, BH 10-m; *92*, BH 3k; *93*, BH 1K; *94*, BH 5k; *95*, BH 9k; *96*, BH 10k; *97*, BH 11k; *98*, BH 10; *99*, Tara, BH 1R; *100*, Novologinovo; *101*, Barabinsk, BH 1R; *102*, BH 146; *103*, BH 303; *104*, BH 23-s; *105*, Kara Sea, Leningradskaya, BH 157.

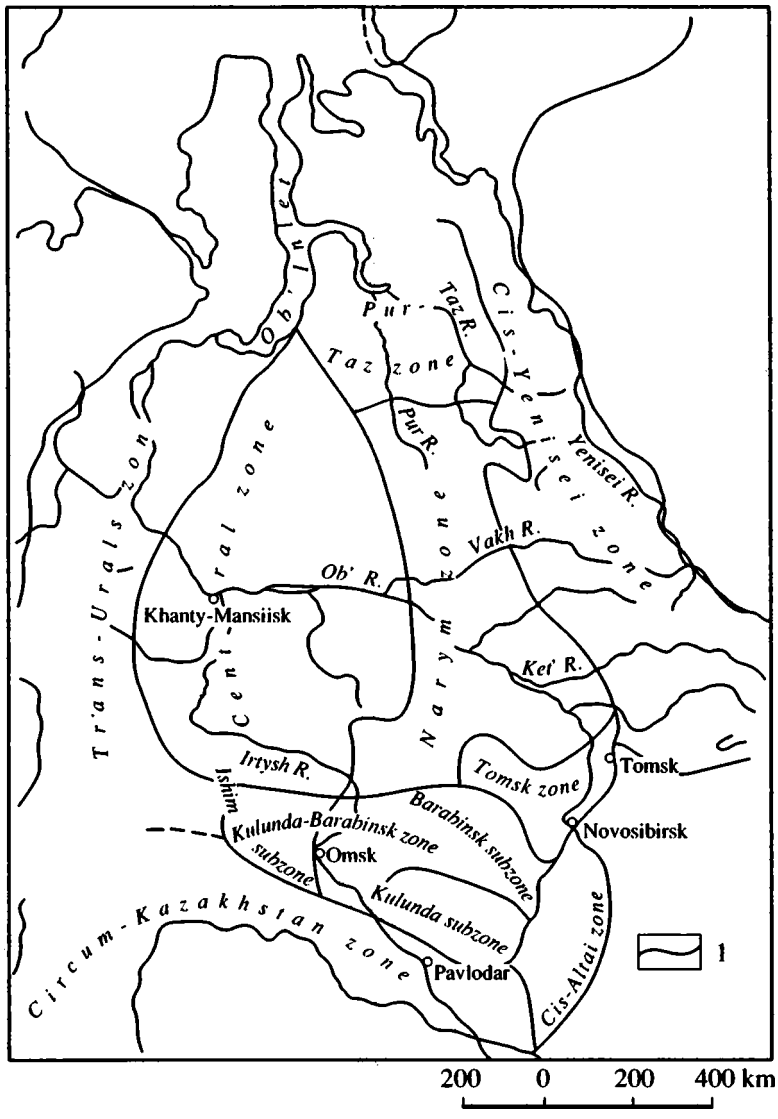


Fig. 3. Structural-facies zones of Paleogene deposits in the West Siberian plate (after Shatskii, 1978): (1) boundary between zones.

Stage coupled with a part of the Thanetian Stage of Western Europe. The Paleocene was divided into the Danian, Montian, and Thanetian stages, which had stratigraphic ranges different from those accepted now (*Stratigrafiya...*, 1975, pp. 45–46; *Paleogenovaya...*, 1975, pp. 286–287). Different age interpretations of the same biostratigraphic units are fraught with serious consequences, first of all with those related to the initially misleading conclusion that the Chegan Formation corresponds in age to the Oligocene and has the diachronous lower boundary. The correct age interpretation of this stratigraphic unit is of particular importance, because everything concerning it concerns as well the Tavda Formation of similar lithology and stratigraphic range.

Speculating on isochronism or diachronism of stratigraphic boundaries between subdivisions of the marine Paleogene in West Siberia, we actually consider the problem of correspondence between bio- and lithostratigraphic units. At present, when we know that open sea basins of the Paleoarctis and Tethys significantly influenced, concurrently or in turn, the biota and sedimentation in the West Siberian epicontinental basin, it seems quite natural to assume diachronism of lithostratigraphic boundaries. First of all, this should be typical of the marginal plate zones, where marine facies grade into continental deposits. Such a situation was recognized long ago by Siberian geologists (Martynov, 1973, pp. 53–54). Accordingly, especially by documentation of the recovered core sections, the same core intervals had been attributed to different formations, boundaries and stratigraphic ranges of which thus received an ambiguous interpretation. We deem it expedient to mention here two demonstrative examples.

(1) In the Ket'–Tym' interfluve area, the southeastern part of the West Siberian plate, Martynov (1973) and Nagorskaya *et al.* (1978) attributed the green clay member overlying the Lyulin-Vor Formation to the lower Tavda Subformation. Podobina (1975) and, afterward, Dargevich and Lysenko (1984) included this member into the Lyulin-Vor Formation, whereas Shatskii (1978) identified it as the Nyurol'ka Formation (see in Dargevich and Lysenko, 1984, p. 80).

(2) Kozlova and Zaionts (1987) who documented the core section of the Borehole "K-14" (or "K") drilled in the Komsomol'skaya area near the Khanty-Mansiisk have recognized the lower, middle and upper Lyulin-Vor subformations within the depth intervals of 300.4–265.4, 265.4–186.4, and 186.4–164.5 m, respectively, and the overlying clay beds were attributed in this case to the Tavda Formation. According to documentation of P.P. Generalov (see in Kul'kova, 1994), the indicated intervals of the lower and middle Lyulin-Vor subformations correspond to the Serov and Irbit formations, and the Nyurol'ka Formation is detected in the depth interval of 186.4–140.0 m. One can easily see that the Nyurol'ka Formation in interpretation of Generalov and Kul'kova corresponds, in terms of Kozlova and Zaionts, to the upper Lyulin-Vor Subformation coupled with a part of the Tavda Formation. Solution of this problem is of key significance, because Kozlova and Zaionts discovered abundant radiolarians *Stylosphaerella irinae* and *Tripodiscinus tumulosus* in the lower beds of the Tavda Formation, which are placed at the Nyurol'ka level by Generalov and Kul'kova. In the same core section, Podobina (1998, p. 60) attributed the depth interval of 170.0–158.0 m to the upper Irbit Subformation corresponding, in her opinion, to the Priabonian. On the other hand, Kul'kova who studied dinocysts ascribed the same interval to the *Charlesdowniea coleothrypta* s.l. Zone DP7 spanning the late Ypresian and the lower half of the Lutetian.

AGE OF MAIN BIOSTRATIGRAPHIC UNITS OF THE MARINE PALEOGENE IN WEST SIBERIA

Cretaceous–Paleogene Boundary Interval

The most important problem of regional biostratigraphy in West Siberia is related, like it was before, to the legitimacy of attributing the upper beds of the Gan'kino Formation, which are widespread everywhere in the plate, to the Danian. In our solution of the problem, we take into account the inferable age of the beds proper, as they yield benthic foraminifers of the *Brotzenella praeacuta* Zone (*Foraminifery...*, 1964), and also the indubitable Danian age of the dinocyst, radiolarian, diatom, silicoflagellate, and foraminiferal assemblages (data on the latter are reported in Amon, 1990), which have been distinguished in many sections of the lower Talitsa Subformation of the trans-Urals and central West Siberia. Similarly significant are the currently accepted stratigraphic range of the Danian Stage and its position in the general scale, as well as the regional hiatus at the top of the Maastrichtian that is well traceable throughout the southern East European platform, in the north of the Turan plate, and within the Turgai depression.

It used to be a common practice to attribute the greater part of the Gan'kino Formation to the Maastrichtian (the *Spiroplectammina kasanzevi* Zone), but the age of its upper horizons remained disputable for half a century at least. Lipman *et al.* (1960) and V.T. Balakhmatova believed that the formation age is limited by the Maastrichtian, whereas the Danian interval corresponds to the lower Talitsa Subformation. Subbotina and Kisel'man were first to note some difference between assemblages of benthic foraminifers from the lower greater part of the formation and from its upper beds. Later, these beds spanning the range of the *Brotzenella praeacuta* Zone were attributed to the Danian, although Subbotina did not consider personally this taxon as critical for the age determination of enclosing rocks (see in *Foraminifery...*, 1964, p. 61). She grounded the above conclusion on Danian planktonic foraminifers *Globanomalina compressa compressa* (Plumm.) and *Parasubbotina pseudobulloides* (Plumm.),[1] which were found by Kipriyanova (1960), and afterward by Kisel'man (1978) in several borehole sections (Shadrinskaya-3 included) recovered in the eastern slope of the Urals. Kipriyanova thought that this plankton characterizes upper horizons of the Gan'kino Formation, but actually, as is established now (Amon, 1990), it originates from the Shadrinsk Beds, which underlie or grade into basal facies of the Talitsa Formation. Accordingly, the idea of Subbotina and Kisel'man, who postulated distribution of beds bearing the Danian planktonic foraminifers over the entire plate area, cannot be accepted as valid even for western and southern regions of the plate. As for the plate area located northward of the Yekaterinburg–Tyumen' line, many boreholes drilled here (the reference Turinskaya borehole among them) evidence that the Talitsa Formation overlies various, certainly Upper Cretaceous horizons of the Gan'kino Formation, or the latter is completely eroded (Papulov, 1964; Sigov, 1969).

As the index-species *Brotzenella praeacuta* appears in the east of the European paleogeographic province at the beginning of the late Maastrichtian and represents the key taxon for this stratigraphic interval (Naidin *et al.*, 1984a, 1984b), it cannot be used, in opinion of Amon (1990), as the indubitable indicator of the Danian Stage. According to Amon, who personally investigated about thirty natural and borehole sections in the trans-Urals, northern Turgai depression, and southwestern margin of the West Siberian plate, the Gan'kino Formation is everywhere overlain, with a distinct hiatus, by the Talitsa and Serov formations, or by the Shadrinsk Beds. The thin member of the latter is

[1] Generic names of planktonic foraminifers are given according to *Atlas...*, 1999.

composed of slightly calcareous to carbonate-free clay separated by an erosion surface from both the Gan'kino and overlying Talitsa formations. Boreholes 304 and 306, which are drilled near the eponymous site of carbonaceous mineral springs, and Borehole 49 (southeastern outskirts of the town of Kurgan) demonstrated that sequences of the Shadrinsk Beds or younger deposits begin with the basal gravel-pebbly horizon bearing phosphorite. Foraminiferal assemblages characterizing the beds are of a mixed composition. In addition to foraminifers redeposited from the Gan'kino Formation, some of which are of the early Maastrichtian age, these assemblages include typical Danian taxa, for instance, *Parasubbotina pseudobulloides* (Plumm.) that is characteristic as well of the lower Talitsa Subformation. As is known, *B. praeacuta* coexists with *P. pseudobulloides* within the lower Danian interval of those rare sections (e.g., in the Mangyshlak), where Danian sediments rest straight on the Maastrichtian deposits. Coexistence of both forms in the Shadrinsk Beds clearly points to the early Danian age of host deposits. In the lower Talitsa Subformation (e.g., within the depth interval 185–178 m of Borehole 19 at the left side of Pyshma River and in several other boreholes), *P. pseudobulloides* coexists with a series of typical Danian taxa indicating the corresponding age of the subformation. This important conclusion of Amon is proved by distribution of other biotic groups in the lower Talitsa Subformation. All these observations indicate that the Gan'kino Formation age is limited by the Maastrichtian.

As for the hiatus at the base of the Danian, it is, like in the East European platform, about 1 m.y. long and corresponds to the *Parvularugoglobigerina eugubina* Zone. This hiatus is traceable in all regions of western Eurasia.

Talitsa Subformation (Horizon)

Deposits of the Talitsa Formation (or of synonymous horizon) are widespread in western, central, and southern regions of the plate. Two subdivisions of the formation represent a single cyclothem. The lower subformation accumulated at the stage corresponding to the transgression of Arctic waters in the plate. During the subsequent stage, sedimentation and biota development were under influence of the Tethyan waters. In the area of the Polunochnyi manganese deposit, the formation is divided into the Marsyaty and Ivdel' subformations (or independent formations, as they were ranked sometimes). In the Yamal Peninsula and Pur–Taz interfluve, deposits under consideration laterally grade into shallow-water sediments of the Tibeisale Formation. Regional dinocyst assemblages from the Talitsa Formation reveal compositional similarity with their zonal assemblages characterizing the Danian and Selandian in stratotype localities. In addition, the lower subformation also yields characteristic Danian assemblages of radiolarians, diatoms, and silicoflagellates, which are comparable to those known from the lower Syzran' Subhorizon of the East European platform.

Lower Talitsa Subformation. The dinocyst assemblage of the Danian *Senoniasphaera inornata* Superzone, or of the lower *Carpatella cornuta* and upper *Cerodinium striatum* zones, were discovered by Vasil'eva in the Marsyaty Formation, an age equivalent of the lower Talitsa Subformation, at the Loz'vinskaya Pristan' site, and the same unit recovered by the Borehole 18A (18 km to the east from the town of Kamyshlov) also yielded these fossils. According to Vasil'eva, similar assemblages occur immediately above the phosphorite-bearing horizon that separates Maastrichtian and overlying deposits in borehole sections 91 (Vogulka site of the Alapaevsk district), E1-8 (eastern outskirts of the Elan' village), and in the quarry Ayat-2. Dinocysts of the *S. inornata* s.l. Superzone are known as well from the Shadrinsk Beds (*C. cornuta* Zone) and lower Talitsa Subformation, which were recovered by Borehole LD-1 near the Lipovskoe village. Aleksandrova distinguished the diverse assemblage of dinocysts of the *S. inornata* Superzone in samples of black fat clay collected by R.B. Krapivner from the Talitsa Formation at the right-hand side of the Ob' River opposite the Irtysh River mouth. In addition to the dominant index-species, the assemblage includes *Hystrichosphaeridium tubiferum, Spiniferites* species, *Achomosphaera ramulifera, Palaeotetradinium minusculum, Ceradinium sibiricum, C. dartmoorua, Laciniadinium* sp., *Fibradinium anettoprense*, and *Fromea laevigata*. At least a half of organic-walled phytoplankton is represented by acritarchs and prasinophytes: *Mycrhystridium* sp., *Leiosphaeridia* sp., and *Plambages torulosa*.

Danian radiolarians typical of the *Cromyocarpus (?) ovatus* Beds were described by Grigor'eva (1975) who discovered them in clayey–siliceous deposits of the Marsyaty Formation of the trans-Urals region. Their presence in the lower Talitsa Subformation of the Borehole section 19, where the associated *P. pseudobulloides* is detected, was later proved by Amon (1990). In addition to the index-species, Kozlova includes into the assemblage the following species characterizing the first evolutionary stage of Paleogene radiolarians: *Gondwanaria (?) alata, Lithorobus longus, Lophophaena mugaica*, and *Hexacontium subglobosus*. According to data of Kozlova and Strel'nikova (1984), the *Cromyocarpus (?) ovatus* assemblage coexists with diatoms of the *Trinacria heibergiana* Zone (*Trinacria heibergiana–Hemiaulus rossicus* Zone according to nomenclature of Strel'nikova) in the Borehole 19-V (depth interval of 198–175 m) of the Ust'-Man'ya site at the Severnaya Sos'va River. The same diatom zone and the *Corbisema (Dictyocha) hostata* Zone of silicoflagellate zonation was established by Glezer in the Marsyaty Formation penetrated by a series of boreholes in the trans-Urals and near the Loz'vinskaya Pristan', where Vasil'eva recognized the dinocysts *Senoniasphaera inornata* Superzone. Spe-

cies *Hemiaulus rossicus* and *Trinacria caudate* are characteristic among diatoms, whereas *Corbisema archangelskiana* and *C. inermis* are remarkable among silicoflagellates. Panova correlated the *Cibicides incognitos* Beds with the *Trudopollis menneri–Nudopollis endangulatus–Oculopollis giganteus* Palynozone (*Prakticheskaya...*, 1990).

All data cited above clearly suggest the Danian age for the greater part of the lower Talitsa Subformation. At the same time, it seems that its accumulation was asynchronous in different regions, where clayey-siliceous deposits were formed concurrently with slightly calcareous or carbonate-free clay of the Shadrinsk Beds. We met an analogous situation analyzing data on Danian deposits in the Lower Volga region and concluded, as a result, that basal siliceous beds of the Syzran' Group are coeval to the Belogorodnya Beds of different lithological composition. In this case, the age control of siliceous and organic-walled phytoplankton assemblages characterizing the same biozones, like in West Siberia, was additionally substantiated by nannoplankton.

Upper beds of the lower Talitsa Subhorizon may be Selandian in age. In four borehole sections recovered near the Kharasavei Settlement, the southwestern Yamal Peninsula, Paleocene deposits of the Tibeisale Formation overlie, according to data of Bugrova, the erosion surface of clayey to silty beds of the Maastrichtian Gan'kino Formation (*Spiroplectammina kasanzevi* Zone). The basal Paleocene horizon of glauconite- and phosphorite-bearing sand grades upward into clayey-silty rocks of the lower Tibeisale (= lower Talitsa) Subformation. The lower 1.5-m-thick interval of the section is barren of foraminifers, but immediately above, there was discovered foraminiferal fauna characteristic of the Selandian. As Bugrova reported, about 70% of foraminiferal species from the Yamal Peninsula were described by F. Brotzen from Selandian deposits of southern Sweden. The fauna-bearing part of the section is distinguished as the *Ceratolamarkina tuberculata* Beds, where this index-species coexists with *Ceratobulimina perplexa, Ammoscalaria friabilis, Cyclammina coksuvorovae, Pseudopolymorphina geieri angusta, Sigmomorphina soluta*, and some other forms. Deposits with a similar but less diverse assemblage are also known in the Pur–Taz interfluve, where they correspond to a part of the Selandian *Ammoscalaria friabilis* Zone of a wider stratigraphic range.

Upper Talitsa Subformation. The subformation is composed predominantly of clay, and the content of siliceous components gradually decreases in its rocks by transition from the underlying lower subformation. Since connections between the West Siberian basin and World Ocean experienced transformations during the Talitsa time, we suspect partial erosion and local hiatuses at the base of the upper Talitsa Subformation. In its greater part, the subformation corresponds to the Selandian and, possibly, to the initial Thanetian, because it yields characteristic dinocysts of the *Alterbidinium circulum–Cerodinium speciosum* Zone known at this level in northwestern Europe (Stouge *et al.*, 2000), foraminifers of the *Ammoscalaria friabilis* and *Cibicidoides favorabilis* zones (according to Podobina, correlation with the latter can be complete or partial), and palynomorphs of the *Trudopollis menneri–Triatriopollenites arboratus–Anacolosidites insignis* Palynozone (nomenclature of Panova). Siliceous microfossils from the upper Talitsa Subformation are unknown. This may be a consequence of limited or even ceased connections between the West Siberian and Arctic basins, the more so as sediments of the upper Tibeisale Subformation are of the coastal marine and continental genesis (Chirva and Lyubomirova, 1973; Shatskii, 1978). As is emphasized in one recent publication (Semochlina *et al.*, 1994), the upper Tibeisale Subformation penetrated by a series of boreholes in the Yamal (1, 8, 9, 12rr, and 12rg) and Tazovskii peninsulas (41, 42, and 49) yields only spores and pollen, but no marine organic remains.

Lyulin-Vor Formation (Horizon)

The second cycle of Paleogene sedimentation was longest one and corresponded to the maximum expansion of sea basin that occupied two thirds of the plate area. It was a time of most intense silica accumulation in the middle and high latitudes of northern hemisphere. In West Siberia, siliceous sedimentation took place not only in the quasiupwelling[2] zone of the eastern trans-Urals, but also outside it in response to tightened connections between the Arctic Ocean and marginal sea basins of the Tethys. Despite some unsolved nomenclature problems, the majority of specialists recognize three stages of sedimentation period under consideration. In the west, at the site of the quasi-upwelling zone, the stages correspond to successive accumulation of the Serov (opokas) and Irbit (diatomites) formations overlain by siliceous-clayey sequence that used to be considered either as a part (upper subformation) of the Irbit Formation (s.l.), or as the independent Nyurol'ka Formation, the age analogue of the upper Lyulin-Vor Subformation. Eastward, in central, eastern and southern structural-facies zones of the plate, the prevailing clayey-siliceous sedimentation resulted in accumulation of the Lyulin-Vor Formation divided into three subformation, which can be reliably correlated with lithostratigraphic subdivisions distinguished in western areas. Three evolutionary stages are traceable as well in the history of West Siberian marine biota. They correspond in general to sedimentation cycles, and the Lyulin-Vor Horizon is subdivided into three subhorizons. Thus, despite a certain vagueness of lithostratigraphic boundaries, there is general correspondence between litho- and biostratigraphic units. Evaluating

[2] Prefix quasi is used to emphasize that, in contrast to classical upwelling zones near the ocean–continent boundary, this one was untypical.

the latter from the standpoint of paleogeographic interrelations, we may state that two initial stages of sedimentation developed under strong influence of Arctic basins, whereas biota and sedimentary processes of the third stage were affected by northern marginal basins of the Tethys.

Lower Lyulin-Vor Subformation (= Serov Formation)

According to the general opinion, the indicated units conformably overlie the Talitsa Formation, though their sections infrequently begin with thin basal beds of glauconite sandstone with siliceous matrix. These beds were deposited at the early stage of transgression that expanded from the north.

The lower Lyulin-Vor Subformation and Serov Formation span the greater part of the Thanetian Stage. Lower parts of both units correspond to the dinocyst *Alisocysta margarita* Zone. *Cerodinium speciosum* that appeared already in the Selandian is another characteristic taxon of this zone, especially in its basal interval. The *Alisocysta margarita* Zone was originally distinguished in the Thanetian of northwestern Europe, where, namely in the North Sea basin, it is quite confidently correlated (Mudge and Bujak, 1996; Moorkens *et al.*, 2000) with nannoplankton zones NP6 (partially), NP7, and NP8 (partially). As for the *A. margarita* Zone equivalents in radiolarian, diatom, and silicoflagellate zonations of West Siberia, these are *Tripodiscinus sengilensis*, *Trinacria ventriculosa–Sheshukovia mirabilis*, and *Corbisema lamellifera–Naviculopsis robusta* zones, respectively. In their upper parts both lithostratigraphic subdivisions correspond to the dinocyst *Apectodinium hyperacanthum* and *Apectodinium augustum* zones correlated with the nannoplankton zone NP9 in Dania, Great Britain, and Belgium (Moorkens *et al.*, 2000). Vasil'eva and Yakovleva traced these zones in many sections of the northern Turgai depression and western areas of the plate (Vasil'eva, 2000; Radionova *et al.*, 2000). In addition to index fossils, these zones yield other species of the genus *Apectodinium* (*A. parvum* locally considered as zonal taxon instead of *A. angustum*, *A. paniculatum*, and some others) in association with *Cerodinium* species, *Glaphyrocysta ordinata*, *Deflandrea dissolute*, and other forms. The *Apectodinium homomorphum* and *A. augustum* zones (actually, the "acme zone" of the genus) are to some degree equivalents of the regional *Pelatospyrus faveolate* Zone of radiolarians and of the *Hemiaulus proteus–Coscinodiscus uralensis* Zone of diatoms, though they both partially correspond to the lower Ypresian acme zone of *Deflandrea oebisfeldensis*. According to Glezer, the mentioned diatom zones correspond to *Dictyocha deflandrei deflandrei* and *D. frenguellii carentis* zones of silicoflagellate zonation.

Index species of the lower zone coexist with the following characteristic taxa: radiolarians *Discopyle (?) totshilinae*, *Clathrocyclas meta*, and *Pelatospyris senta*; diatoms *Podosira (= Coscinodiscus) anissimovae*, *Grunowiella gemmata*, and others; silicoflagellates *Dictyocha frenguellii* and *Naviculopsis constricta*. Associated taxa from upper zones are radiolarians *Clathrocyclas lipmanae*, *Theocalyptra Aurelia*, and *Lychnocanium zhamoidae*, diatoms *Pyxidicula moelleri* and *Stephanopyxis edita*, and also silicoflagellates *Corbisema lamellifera*, *Dictyocha frenguellii*, and *Naviculopsis aspera*.

Coming now to the boundary between the lower and middle Lyulin-Vor subhorizons, we have to discuss first the problem of the Paleocene–Eocene boundary.

Problem of the Paleocene–Eocene Boundary

Like in other region of the world, the problem of Paleocene–Eocene boundary cannot be easily solved in West Siberia. Even the specialized Working Group that was headed during eight years by M. Aubry and organized by the Subcommission on Paleogene System of the International Stratigraphic Commission (ISC) was unable to find an ultimate solution (Aubry *et al.*, 1999). Aubry also guided the IGCP project 308, the main purpose of which was to analyze geological and biotic events across the Paleocene–Eocene boundary.

Members of the international subcommission considered the following possibilities:

(1) To place the boundary in question at the level separating the nannoplankton subzones NP10a and NP10b, and also corresponding to the FAD of *Tribrachiatus digitalis*; this level accepted now by many biostratigraphers coincides almost precisely with the London Clay base and is 54.37 Ma old according to isotopic age determination (Aubry, 1996).

(2) To consider the boundary as separating zones of planktonic foraminifers P5/P6 and corresponding to the LAD of *Morozovella velascoensis*; in Russia, this viewpoint is shared by V.A. Krasheninnikov.

(3) To place the boundary between the nannoplankton zones NP9 and NP10 like it was done by the ISC Commission of Paleogene System in 1989 and authorized by the ISC.

(4) To position the boundary at the level of $\delta^{13}C$ excursion dated back to 55.5 Ma (Kennett and Scott, 1981).

(5) To correlate it with the magnetic polarity inversion between chrons C25n/C24n.

(6) To consider the top of Thanetian sands as corresponding to the desired boundary. The time difference between extreme versions is over 2 m.y. (Aubry *et al.*, 1999). The International Working Group concluded that versions 1 and 4 are most appropriate. The first one having historical grounds is widely accepted in the world as the "standard" position of boundary between the Thanetian and Ypresian stages, which is, however, hardly traceable over the globe. Next version of the so-called "Global Boundary Stratotype Section and Point" is convenient for the worldwide correlation.

In West Siberia, where planktonic foraminifers and nannoplankton from the boundary interval, as well as data on C-isotope excursion, are unknown, we should use, when defining the boundary position, some other criteria, e.g., distribution of another microplankton group. Dinocysts appear to be more suitable than other fossils, because their distribution is less dependent on the facies criterion. In addition, compositional differences between West Siberian and West European dinocyst assemblages are not very significant, and they can be correlated quite reliably. Especially interesting in this respect is the acme zone of *Deflandrea oebisfeldensis* that separates the *Apectodinium augustum* (Paleocene) and *Wetzeliella meckelfeldensis meckelfeldensis* (Eocene) zones in West Siberian sections. *D. oebisfeldensis* that appears in the Paleocene becomes abundant in basal beds of the Serov Formation and middle Lyulin-Vor Subformation, and the beds' base is in fact the lower boundary of the corresponding acme zone. Studying borehole sections in the type area of the Ypresian Stage in Belgium, Moorkens *et al.* (2000) established that the acme zone under consideration spans the time range of 54.8–54.5 Ma. The mass abundance of *D. oebisfeldensis* in the North Sea is recorded at the same datum (Powell, 1988; Heilmann-Clausen, 1994; Mudge and Bujak, 1994, 1996). The datum corresponds to the NP10 zone of the nannoplankton scale. Like in West European sections, upper beds of the West Siberian Serov Formation yield abundant representatives of the genus *Apectodinium* (*A. homomorphum, A. hyperacanthum*, and *A. augustum*) and represent a kind of the *Apectodinium* spp. acme zone. Boundary between the *Apectodinium* and *Deflandrea oebisfeldensis* acme zones is easily detectable in West Siberia. It separates here lithostratigraphic subdivisions and, consequently, may represent the desired boundary. Relationships between the *D. oebisfeldensis* acme zone and other zones of siliceous microplankton are established in the quarry section not far away from the town of Kamyshlov. Concurrent upper boundaries of the *Coscinodiscus uralensis* and *Pelatospyris foveolata* zones are located in the upper part of the acme zone, at the level 1 m below the FAD of *Wetzeliella meckelfeldensis* in the lower part of the Irbit Formation.

Another option for West Siberia is to place the Paleocene–Eocene boundary at the base of the *Wetzeliella meckelfeldensis* Zone (= upper boundary of the *D. oebisfeldensis* acme zone). In this case, we omit however the lowermost *W. astra* Zone of the Ypresian dinocyst zonation. The last zone was so far established by Kul'kova in a single West Siberian section, where its stratigraphic range is very short. An advantage of this version is that the boundary will be precisely between the *Pelatospyris foveolata* and *P. fiscella* zones of radiolarians, and simultaneously, between the *Coscinodiscus uralensis* and *C. payeri* zones of diatoms. At the same time, its position will be, unfortunately, not between lithostratigraphic units, but inside one of them.

Middle and Upper Lyulin-Vor Subformations (Irbit and Nyurol'ka Formations)

The first half of the Eocene corresponded to the accumulation time of siliceous sediments (diatomites) replaced in the lateral direction (from the west to the east) and upward in the section by clayey-siliceous facies. These sediments overlie older lithostratigraphic units without a hiatus. The complete succession of four Ypresian dinocyst zones (*Wetzeliella meckelfeldensis–Dracodinium simile–Dracodinium varielongitudum–Charlesdowniea coleothrypta* s.l.) is traceable in West Siberia, like in Western Europe and Crimea–Caucasus region, above the *D. oebisfeldensis* acme zone. Three lower dinocyst zones correspond in their summary range to nannoplankton zones NP11 and NP12. The fourth zone is an equivalent of nannoplankton zones NP13 and NP14 (presumably coupled with a part of the NP15 zone). The distinguished regional zones of dinocysts are well correlative with zonal subdivisions of other microplankton groups (Fig. 1), as it is established in a series of sections studied in the quarry near the town of Kamyshlov and recovered by boreholes in the trans-Urals, Pur–Taz, and Central structural-facies zones (our own and additional data of Siberian experts in palynology; see Kul'kova and Shatskii, 1984; Bakieva, 1994; Kul'kova, 1994).

Relationships between the Ypresian organic-walled and siliceous microplankton zonations are as follows: the interval, incorporating a part of the *Deflandrea oebisfeldensis* acme zone together with the *Wetzeliella meckelfeldensis* and *Dracodinium simile* zones coupled with a part of the *Dracodinium varielongitudum* Zone, corresponds to the lower Ypresian *Pelatospyris fiscella* and *Podocyrtis aphorma* zones of radiolarian scale, to the *Coscinodiscus payeri* Zone of diatoms, and to the *Dictyocha frenguellii carentis* Zone of silicoflagellates. In opinion of Glezer, the latter better corresponds to the *Deflandrea oebisfeldensis* acme zone. Podobina (1998) attributed the *Textularia carinatiformis sibirica–Anomalinoides ypresensis ovatus* Zone to the Ypresian Stage, but in our opinion, it corresponds to three lower zones of the Ypresian dinocyst zonation and spans only a part of the *Dracodinium varielongitudum* Zone. The greater upper part of the last zone terminates the lower half of the Ypresian and corresponds to the lower subunit of the radiolarian *Heliodiscus lentis* s.l. Zone subdivided by Kozlova into the *H. lentis* s.s. and *H. inca* subzones. The dinocyst zone in question is an equivalent of the *Pyxilla gracilis* (nomenclature of Strel'nikova, Radionova and Oreshkina) or *Coscinodiscus decrescens decrescens* (nomenclature of Glezer) zones of the diatom scale. The correlative unit of silicoflagellate zonation is the *Dictyocha rotundata* Zone (in terms of Glezer). Biostratigraphic units listed above characterize the Irbit Formation (s. str.) and the middle Lyulin-Vor Subformation.

The following characteristic taxa of siliceous plankton were detected in the considered biostratigraphic

units: *Clathrocyclas elegans, C. multiplicatus* and *Thecosphaerella sublica* from the *Pelatospyris fiscella–Axoprinum inclarum* Zone, as well as *Podospirites papalis* and *Lamptonium fabaeforme fabaeforme* from the *Podocyrtis aphorma* (= *Spongotrochus paciferus*)–*Axoprinum mirum* Zone of radiolarians; *Coscinodiscus mirabilis, C. heteroporus* and some others from the *Coscinodiscus payeri–C. decrescens* Zone of diatoms; and silicoflagellate *Dictyocha deflandrei* var. *completa.* According to Panova, the main part of the *Pelatospyris fiscella* Zone that was distinguished in the Borehole "K" (depth interval of 236.5–210.0 m, Komsomol'skaya area) yields palynological assemblage of the *Triporopollenites robustus–Plicapollis pseudoexcelsus* Zone. As for the *Castanea crenataeformis–Platycaryapollis* Palynozone (depth interval of 210.0–172.0 m), Kul'kova correlated it with the dinocyst *Dracodinium varielongitudum* Zone coupled with the lower part of the *Charlesdowniea coleothrypta* s.l. Zone.

The upper half of the Ypresian and, presumably, a part of the Lutetian are correlative with the dinocyst *Charlesdowniea coleothrypta* s.l. Zone that is used to subdivided into the *Charlesdowniea coleothrypta* s.s. and *Charlesdowniea coleothrypta* sbsp. *rotundata* subzones. In a wide sense, this zone spans upper beds of the Irbit Formation (= middle Lyulin-Vor Subformation) and the whole (or partial) range of the Nyurol'ka Formation or upper Lyulin-Vor Subformation.

The *Charlesdowniea coleothrypta* s.s. Subzone corresponds to the radiolarian *Heliodiscus inca* Subzone (upper one in the *Heliodiscus lentis* s.l. Zone). One more *Buryella clinata–Buryella longa* Zone of West Siberia is tentatively distinguished in the upper Lyulin-Vor Subformation penetrated by Borehole 011-BP near the Achair Settlement (depth interval of 485–470 m). The *Charlesdowniea coleothrypta* s.s. Zone spans an upper part of the diatom *Pyxilla gracilis* Zone and adjacent part of the overlying *Pyxilla oligocaenica* var. *tenius* Zone. According to Strel'nikova, Oreshkina, and Radionova, the latter is uppermost in the Ypresian Stage. Glezer considers the upper part of the *Pyxilla gracilis* Zone as the independent *Brightwellia hyperborean* Zone and attributes the *Pyxilla oligocaenica* var. *tenius* Zone to the Eocene without determination of its upper limitation (Glezer, 1979). In addition, the *Charlesdowniea coleothrypta* s.s. Subzone is correlative with the *Dictyocha secta* Zone of silicoflagellates, and we believe that it is still of the Ypresian age, whereas the *Charlesdowniea coleothrypta rotundata* Subzone spans the late Ypresian–Lutetian interval. The upper limitation of the latter should be specified, like that of the correlative (wholly or partially) *Pyxilla oligocaenica* var, *tenius* Zone of diatoms and *Lychnocanium separatum* Zone of radiolarians. We suggest that all these subdivisions are concurrent with or closely correspond to the stage of almost complete cessation of silica accumulation in central areas of West Siberia that happened in the mid-Lutetian time. It is unclear, however, whether the *Pyxilla oligocaenica* var. *tenius* Zone extends up to lower beds of the Tavda Formation, or not. The same problem is urgent in regard to the *Lychnocanium (?) separatum* Beds distinguished by Kozlova (1999).

Relationships between the Lyulin-Vor and Tavda Formations

Considering lithostratigraphic subdivisions, we should emphasize that there are contrary interpretations of geological relations between the Lyulin-Vor (Nyurol'ka) and Tavda formations, and this important problem of the marine Paleogene stratigraphy in West Siberia still cannot find an adequate solution. According to one of contrary opinions, the Tavda Formation overlies the Lyulin-Vor sequence with a hiatus everywhere (Shatskii, 1984, pp. 12–13). In contrast, Podobina (1998) argued for the gradual transition between both formations pointing out that there was no break in sedimentation.

Examples convincing that the Tavda Formation begins with a basal horizon of shingle or glauconite sand bearing phosphorite nodules are presented in many publications and unpublished reports of mapping geologists, where this situation is interpreted as indicating a hiatus in sedimentation. Lithostratigraphic relationships of this kind are established in the trans-Urals and also in some northern, southern, and central area of the plate, where the lower Tavda beds of sand rest on the erosion surface of underlying deposits. Vlasov and Cherepanov (1984, p. 82) observed these relationships in the Irtysh River lower courses, and Kriventsov (1984, p. 90, Fig. 2) presented similar data on the eastern Ob'–Irtysh interfluve area, where the erosion surface at the base of Tavda Formation was recognized in borehole sections 10i, 1k, 4k, 5k, and 45p. The described viewpoint was shared to some or other extent by Martynov (1973), Chirva and Lyubomirova (1973), and also by Shatskii (1984) who argued that the erosion surface between the Lyulin-Vor and Tavda formation is occurring everywhere (see his scheme in pp. 12–13). Quite typical relationships between these formations can be observed in core section recovered by Borehole 011-BP near the Achair Settlement of the Omsk district. At the contact between clayey and silty to sandy beds, there are "hardground" marks, above which basal sediments of the Tavda Formation are bioturbated and show signs of suspended load deposition. The contact zone is recorded by specific pattern deviations in the down-hole logging diagram. Kul'kova who studied dinocysts from this core section established that their assemblage characterizes the *Rhombodinium draco–Areosphaeridium dictyoplokus* Zone. The analogous zone named after the firs species was distinguished at the base of the Bartonian type section in England (Powell, 1992). Since upper beds of the Lyulin-Vor Formation still correspond to the *Charlesdowniea coleothrypta* s.l. Zone, at least two dinocyst zones (*Charlesdowniea fasciata* and *Systematophora placacantha*)

appear to be missing from the section under consideration.

The last conclusion is supported by Vasil'eva and Yakovleva who studied dinocysts from the upper Lyulin-Vor Subformation and Tavda Formation. The above example is not a single one demonstrating gap that corresponds to one or two dinocyst zones of the terminal Lutetian (?)–initial Bartonian interval. This is clearly manifested in visible exposures, e.g., in the Kamyshlov, Kachar and other quarries, sections of which can be investigated in detail. We suspect that loose and thin beds of glauconite sand, gravel and shingle, which mark the contact under discussion, used to be washed out in the course of drilling, because special efforts to increase the core recovery (e.g., double-barreled core drilling with frequent sampling) have not been undertaken in practice and correct inferences are possible only from borehole logging diagrams, when they are available. Some of us had such an experience in Kazakhstan.

In those structural-facies zones, where the transition from the upper Lyulin-Vor Subformation to Tavda Formation is interpreted as gradual, some researchers distinguish a member of the "Chegan Clay." This uncertainly recognizable lithostratigraphic unit is attributed to the upper Lyulin-Vor Subformation in some cases and to the Tavda Formation in the others. The dinocyst assemblage from the member is considered as characterizing a particular zone variably named after one or another species occurring here, for instance, after *Systematophora placacantha* and *Charlesdowniea fasciata* or, occasionally, after *Wetzeliella articulata* and some other species. Despite of all nomenclature uncertainties, the zone is easily distinguishable because of predominance of *Wetzeliella* and *Deflandrea* forms. In the dinocyst zonation, this zone is transitional between the *Charlesdowniea coleothrypta* sbsp. *rotundata* Subzone and *Rhombodinium draco* Zone undoubtedly characterizing the Nyurol'ka and Tavda formations, respectively. The underlying subzone is distinctly not younger than the initial Lutetian. As for the upper zone, it is basal one in the Bartonian biozonation and correlative with the *Subbotina turcmenica* Zone known in western and central Europe, Crimea–Caucasus region, and in the south of the East European platform.

Analyzing five samples from the depth levels of 444, 439, 434, 424, and 419 m, Zaporozhets distinguished the zone under discussion in the borehole section 2 drilled in the Kaimysovy Arch, the Vasyugan River basin. According to Podobina (1998), the section part between the *coleothrypta* and *draco* zones belongs to the upper Lyulin-Vor Subformation penetrated here within the depth interval of 446–407 m. The association of abundant and diverse phytoplankton (over 50 species) with infrequent spores and pollen suggests that their host deposits accumulated in an open marine basin with normal salinity. The generic diversity of the assemblage is high. In addition to *Wetzeliella* and *Deflandrea* forms representing more than a half of the assemblage, it includes *Areoligera coronata, Thalassiphora pelagica, T. delicata, Diacrocanthidium spinigerum, Samlandia chlamidophora, Soaniella granulata, Achomosphaera alcicornu, Cordosphaeridium gracile, Homotryblium acculeatum, Lejeunecysta wetzelii, Rottnestia borussica, Spiniferites cornutus*, and *Cribroperidinium guiseppei*, which associate with 20 other taxa classified in the open nomenclature. The lower age limit of the assemblage is defined quite precisely. In the core section, beginning from the depth level of 449 m, there is established the following descending succession of biozones: *Charlesdowniea coleothrypta* (s.l.)–*Dracodinium varielongitudum* (474 m)–*Wetzeliella meckelfeldensis* (489 m). The upper boundary of the assemblage distribution is undetected, but we should mention that *Ch. coleothrypta* and *Ch. coleothrypta* sbsp. *rotundata* are quite abundant in the sample from the depth level of 419 m, and this may indicate that the assemblage characterizes an upper part of the *Ch. coleothrypta* s.l. Zone. The comparable results were obtained by Kul'kova and Yakovleva who studied dinocysts, spores and pollen from the borehole section 4 drilled nearby in the same arch. The only difference is that the upper Lyulin-Vor Subformation (depth interval of 310–252.9 m) is correlated by these researchers with the *Ch. coleothrypta rotundata* s.l. Zone. The appearance of the genus *Soaniella* typical of the "Chegan Clay" is recorded in this case at the depth level of 256 m, i.e., in the uppermost part of the subformation.

In the borehole section 2, Podobina (1998, p. 87) selected the depth interval of 430–419 m for the stratotype of the newly distinguished *Labospyra honesta* Zone of the Priabonian age, as she believed. Accordingly, she attributed the upper Lyulin-Vor Subformation to the upper Eocene. In the borehole section 4, Podobina (1998, p. 74) recognized the same zone in the depth interval of 280–270 m.

Thus, as one can see, experts on different fossil groups suggest quite different ages for the upper Lyulin-Vor Subformation. In accordance with distribution of organic-walled phytoplankton, it appears to be not younger that the Lutetian, whereas the late Eocene age is inferred on the basis of foraminifers. Such a great discrepancy in age determination is a problem of special consideration.

Tavda Formation (Horizon)

The third Paleogene cycle of marine sedimentation resulted in accumulation of the Tavda Formation of finely laminated clay beds, thickness of which is up to 200 m. Together with concurrent continental facies, these beds represent the Tavda Horizon. A bed of glauconite sand widespread in the region and best manifested in southern structural-facies zones separates the lower and upper Tavda subformations. The accumulation time of sediments under consideration corre-

sponded to the terminal stage of the Paleogene marine sedimentation. Siliceous phytoplankton is either missing from the formation, or occurs in its basal beds, but it is unclear, however, if this group of foffils is allochthonous or buried "*in situ*." The Tavda Formation age is interpreted controversially. According to one viewpoint, it is of the middle–late Eocene age, but other researchers attribute this unit to the Oligocene. Irrespective of that or other viewpoint, one should take into account, when analyzing the corresponding age problem, the distribution of foraminifers (benthic and planktonic), dinocysts, ostracodes, higher plant remains, carpolites, fishes, and mollusks.

Benthic and planktonic foraminifers. Podobina argued for the Oligocene age of the Tavda Formation in her many publications, the monograph of 1998 included. Her main argument is occurrence of species *Evolutononion decoratum* in the *Cibicidoides pseudoungerianus* Zone. The section regarded by Podobina as lectostratotype of the distinguished zone is that penetrated in the depth interval of 115.75–37.75 m by Borehole 4-k near the village of Petukhovo in the Omsk district. Description of this section is well known from an early publication (*Foraminifery...*, 1964). In opinion of Podobina, the zone yields Oligocene foraminifers and is identical in range to the previously distinguished *Cibicides khanabadensis* (nomenclature of Lipman) or *Elphidium rishtanicum–Cibicides khanabadensis* (nomenclature of Ushakova) zones, and also to the *Cribrononion* (= *Criptoperidinium rishtanicum*) Beds (*Foraminifery...*, 1964). The last taxon has been described from the Rishtan Beds of Central Asia and represents now the index species of the upper local zone of the Bartonian Stage in southeastern Turkmenistan and Tajikistan (an equivalent of the upper portion of the *Globigerina turcmenica* Zone).

Selecting the lectostratotype of newly distinguished zone (see above), Podobina did not realize that Subbotina (see in *Foraminifery...*, 1964, pp. 72–73) who studied the same borehole section discovered at the depth level of 50 m, i.e., in upper beds of her lectostratotype, an assemblage of planktonic foraminifers similar in composition to that from the *Lyrolepis* Beds of the northern Caucasus (Kuma Horizon in current interpretation). *Globigerina aperetura* (current synonym *Subbotina turcmenica*) described by Subbotina among other foraminifers is a species convincing that planktonic foraminifers form West Siberia are similar in composition to their assemblage from the Kuma Horizon of the Crimea–Caucasus region, where this unit is dated back to the middle Eocene on the basis of planktonic foraminifers and nannoplankton. Moreover, Subbotina was sure that an analogous assemblage of planktonic foraminifers was detected by Ushakova (1959) in the Ozernyi section of the Kustanai district. Subbotina suggested therewith that the assemblage in question characterizes the upper Nyurol'ka beds but not the Tavda Formation. In our opinion, Podobina more correctly attributes the respective interval of the borehole section 4 to the Tavda Formation.

As for the *C. rischtanicum* Beds, they have been always considered by Subbotina and Ushakova as a member of the Tavda Formation. This taxon, if we can consider it as reliably indicative, is widespread in higher horizons of West Siberian sections, above the beds with planktonic foraminifers. As is known, Ushakova who analyzed distribution of elphidiids in the Tavda Horizon distinguished three *Elphidium* zones each yielding as well *C. rischtanicum*. According to Ushakova, elphidiids are missing from uppermost and basal beds of the third *C. differensapertio* Zone crowning the Tavda Formation section. Podobina (1998, Figs. 25 and 26, pp. 73–74) shared the opinion of Ushakova and Subbotina about stratigraphic position of the *C. differensapertio* assemblage.

In 1999, Beniamovskii identified species *S. turcmenica* among planktonic foraminifers from lower clay beds of the Tavda Formation exposed in the Kamyshlov quarry. The species identification was proved by Bugrova who revised his collection. The beds in question rest on the erosion surface of the Irbit Formation and also overlie a thin (2–3 m) bed of loose sandstone that is separated by erosion surfaces from both the underlying Irbit and overlying Tavda formations. Age of the last bed is unclear, because it is barren of siliceous microfossils and yields only organic-walled phytoplankton of a wide stratigraphic range, namely acritarchs and prasinophytes of the *Cymatiosphaera, Ptanospermella, Tytthodiscus*, and *Tetraporina* genera (determinations of Zaporozhets).

In order to elucidate the Tavda Formation age, it would be purposeful to study foraminifers in the borehole section 001-BP drilled near the Achair Settlement of the Omsk district. Like the Petukhovo and Ozernyi sites, this one is located between the Kulunda–Barabinsk and circum-Kazakhstan structural-facies zones, where the Tavda Formation is thicker than elsewhere. It is important to establish here whether we have one or several datums of planktonic foraminifers in this formation of West Siberia and how correct are the early conclusions of Subbotina (*Foraminifery...*, 1964) from the viewpoint of recently obtained data. Judging from replacement of agglutinated foraminifers by calcareous ones, we suspect that planktonic representatives of the group penetrated in the West Siberian basin via the Turgai trough from the Crimea-Caucasus region and Central Asia. This is evident from benthic species present in the Tavda Formation, which are identical (vicarious) to *Cibicides khanabadensis, Cribrononion rishtanicum*, and some other forms from Central Asia. In the 1950–1960s, the Isfara–Khanabad and Sumsar beds of the Fergana and Tajik depressions were attributed to the Oligocene. At present, when the late Eocene nannoplankton is found to be typical of the Khanabad Formation and also of lower marine sediments of the Sumsar Formation (Muzylev and Salibaev, 1988; Salibaev

et al., 1995), it would be incorrect to ascribe the Oligocene age to the Tavda Formation based on the outdated age interpretation of biostratigraphic units in Central Asia.

All other biotic groups considered below (except for those from the Kurgan Beds) suggest the late Eocene age of the Tavda Formation.

Ostracodes. According to Nikolaeva who investigated a series of borehole sections in the Tyumen', Ivlevo, Pokrovskaya, Tebis (Well 3-k), and Zavodoukovsk (Well 1-k) prospect areas, *Cytheridea probata* Mand., *Echinocythereis spongiosa* (Liepin), *Pterigocythereis permira* Mand., and *Eopaijaenborchella trisulcata* (Mand.) are characteristic ostracode species from the Tavda Formation. Many of them are transient, and typical Oligocene forms are unknown among these taxa. Representatives of the genus *Eopaijaenborchella* do not occur in Oligocene deposits of the East European platform and Turan plate. The majority of listed species is known from the Chegan Formation of Kazakhstan.

Fishes. In western areas of the plate, Elasmobranchii remains occur in lower and upper Tavda subformations (Zhelezko and Kozlov, 1999). Abundant teeth of sharks *Striatolamia macrota* stage *rossica* and *Jaekelotodus trigonalis* from the lower subformation belong to the *Otodus poseidoni ustjurtensis* Zone (E11). The upper subformation corresponds to the *Otodus sokolovi sokolovi* Zone (E15, the middle Priabonian zone in the regional Elasmobranchii zonation of Eurasia) and yields guide fossils of the latter: *Jaekelotodus trigonalis* stage *kacharensis, Striatolamia sibirica*, and *Araloselachus turgaensis* stage *turgaensis*. Similar shark species are typical of the Chegan Formation in the southern Turgai and northern Aral depressions. The first taxon is also known from the Aldaevo Formation of Mangyshlak and from the Obukhov Formation of Ukraine. In the west of the Turan plate, shark teeth characteristic of the selachioid zone E15 occur in beds bearing planktonic foraminifers of the *Globigerapsis tropicalis* Zone and nannoplankton characteristic of the boundary interval between the NP19/NP20 and NP21 zones.

Mollusks. In the 1950–1960s, when the majority of researchers attributed the Tavda and Chegan formations to the Oligocene, N.K. Ovechkin and I.A. Korobkov considered mollusks from these units as corresponding in age to the Eocene. Later, Korobkov (1984) validated his early conclusions. O.V. Amitrov, I.A. Goncharova, and S.V. Popov believe that the molluscan assemblage of *Turritella oksyceras* Eichw., *Sinum rotundatum* (Koenen), *Aporhais* aff. *cornutus* Alexeev, and some other forms is an impoverished analogue of the late Eocene assemblage from the Chegan Formation.

Paleocarpology. Nikintin (1998) distinguished two datums with carpolites in the Tavda Formation: the lower *Azollites minor–Pseudoisoeites tymensis* and upper *Azolla oligocaenica–Alnus cheganica–Decodon sphenosus* zones. The zones were attributed to the Bartonian and Priabonian stages, respectively.

Spores, pollen, and dinocysts. Investigating palynomorphs from the Tavda Formation, many experts in palynology shared the Panova's opinion that the lower Tavda Subformation corresponds to the *Quercus gracilis–Castanopsis pseudocingulum (+ Tricoloropollenites liblarensis)* zone of the Bartonian and, probably, of the initial Priabonian, whereas the upper subformation is correlative with the Priabonian *Quercus gracilis–Quercus graciliformis* Zone. Upper beds of the formation yield abundant spores of fern *Hydropteris indutus* that is an indicator of incipient freshening stage in the Tavda basin evolution. Other freshening episodes are detectable at the lower occurrence levels of buoyant fern *Azolla vera* Krysht. On the basis of identical sampling intervals, Vasil'eva, Zaporozhets, Kul'kova, and Yakovleva correlated palynozones, which were distinguished by Panova in the northern Kazakhstan, trans-Urals region, and West Siberia, with the organic-walled phytoplankton zonation. They established that the lower palynozone spans the interval of dinocyst *Rhombodinium draco* and *Kisselovia ornata* zones corresponding to zones D10 and D11 of the West European standard.

In southern England, the *Rhombodinium draco* Zone is lowermost in the Bartonian type section (Powell, 1992). The group of associated taxa includes *Rhombodinium porosum, Areosphaeridium dictyoplokus*, and some other forms (Vasil'eva, 1994). The *K. ornata* Zone (Vasil'eva, 1994; Kul'kova, 1994) is reliably established in the trans-Urals and Siberia, but its index species is unknown in Paleogene deposits of Western Europe. This zone spans an upper interval of the lower Tavda Subformation, as it is established in borehole sections K-14 (Khanty-Mansiisk district), IG-32, VK-36, VK-49 (Kurgan district) and in some other areas. The index species appears at this level in association with *Wetzeliella irtyshensis* and *Thalassiphora elongata.*

The upper Tavda Subformation corresponds to the dinocyst *Wetzeliella clathrata* sbsp. *angulosa* Zone. This taxon is an index species in the upper Eocene dinocyst zonation of the Crimea–Caucasus region, France, Italy, and central Europe. Its position in the general scale is controlled by coexisting upper Eocene assemblages of planktonic foraminifers and nannoplankton (Andreeva-Grigorovich, 1991; Costa and Manum, 1988; Krasheninnikov *et al.*, 1989; Akhmet'ev and Zaporozhets, 1992). As for the *Quercus gracilis–Quercus graciliformis* Palynozone of the upper Eocene, it is uniformly established in the Crimea–Caucasus region, East European platform, Scythian and Turan plates, Turgai depression, and West Siberia.

Kurgan Beds

The Kurgan Beds, stratigraphic and geographic ranges of which are ambiguous to a certain extent, used to be considered as terminal subdivision of the marine Paleogene sequence. Arguments in favor of distinguishing the beds as an individual bio- and lithostratigraphic unit are as follows:

(1) The pollen assemblage, where dominant components are represented by equally abundant conifers and Fagaceae of the *Quercus gracilis–Quercus graciliformis* group, is peculiar of these beds only. In opinion of Panova (1970), who was first to distinguish the beds as a lithostratigraphic subdivision, this suggests the successive character of the assemblage with respect to the older one characterizing the late Tavda time.

(2) The beds yield an impoverished molluscan assemblage different from that of the Tavda Formation and resembling the early Oligocene assemblage from the Ashcheiaryk Formation of the northern Ust'yurt. Korobkov (1966) was first to note this situation, though sandstones bearing the early Oligocene fossils were known in the trans-Urals after the early works of A.P. Karpinsky in the 1880s.

(3) In their brown, brownish yellow to ashy-gray colors and lithology, the Kurgan Beds resemble deposits of the overlying Kurtamysh and Atlym formations.

(4) The peculiar spore-and-pollen assemblage of the beds coexists with dinocysts and prasinophytes known from basal strata of the lower Oligocene Ashcheiaryk Formation of the Turan plate and from coeval deposits of the Turgai depression.

Especially important in the last case is presence of *Phthanoperidinium amoenum* representing the index species of the lowermost Oligocene zone (D13) in the Crimea-Caucasus region and Western Europe. In West Siberia, Zaporozhets detected the same dinocyst level studying the marine sedimentary sequence recovered by boreholes 1 and 2 within the Kaimysovy arch of the Vasyugan River basin. Unfortunately, Krasheninnikov and Muzylev failed to find foraminifers and nannoplankton in the duplicates of samples studied by Zaporozhets. An analogous dinocyst assemblage was discovered by her in sandstones bearing Oligocene mollusks and exposed in several quarries near the town of Kurgan. Vasil'eva reported that *P. amoenum* occurs in sandstones overlying the Tavda (Chegan) Formation in the Kachar quarry exposure (northern Turgai depression). Not far away from this quarry, in exposures located to the north of the Kushmurun Lake, Korobkov (1966) collected the Ashcheiaryk assemblage of mollusks. According to data of Kul'kova, a species presumably representing *Phthanoperidinium amoenum* was found in a sample collected directly at the contact between the Tavda and Isil'kyul' formations penetrated by Borehole 011-BP near the Omsk.

The geological position of the Kurgan Beds (s.l.) is interpreted controversially. According to one opinion, they accumulated during the sea freshening prior to its final desiccation and represent the uppermost member of the Tavda Formation. The successive character of the Tavda and Kurgan palynoassemblages used to be the main argument in favor of this opinion. Presumable age equivalents of the Kurgan Beds were distinguished as the Aleshkiny Yurty Formation (Nikolaev, 1947), or as Konolevka Member (Zakharov, 1968) in the trans-Urals, Central structural zone, and in middle to lower courses of the Ob' River.

The contrary opinion originally formulated by Korobkov (1966) suggests that the Kurgan beds accumulated during the Oligocene sedimentation cycle. Those of us who participated the expedition of 1999 to the trans-Urals (Aleksandrova, Akhmet'ev, Beniamovskii, Vasil'eva, and Zhelezko) studied relationships between the Tavda Formation and Kurgan beds in quarry sections near the town of Kurgan. We share the opinion of Korobkov, because the Kurgan Beds (s. str.) rest here in the erosional gully that is incised into the Tavda clay partially affected by weathering prior to deposition of the beds. In this case, they may represent marine facies of the Kurtamysh and Atlym formations, which accumulated during seawater ingression into West Siberia via the Turgai strait from the Turan basin. All arguments in favor of this scenario will be considered in a special paper (Akhmet'ev *et al.*, in press), and here we should mention only that the suggested model well explains numerous occurrences of the early Oligocene phytoplankton and mollusks in the Turgai depression and in southern and southwestern areas of the plate. To disprove it, one should convincingly explain how these representatives of marine biota appeared in West Siberia earlier that in other extratropical region of Eurasia.

Concluding our discussion, we should stress that the refined stratigraphy of marine Paleogene deposits in West Siberia is important for age substantiation of coeval continental facies in marginal zones of the plate.

CONNECTIONS BETWEEN THE EPICONTINENTAL WEST SIBERIAN SEA AND PALEOOCEANC BASINS

The Paleogene geological evolution of West Siberia cannot be understood in full measure, if we omit analysis of paleogeographic environments during three principal stages of marine sedimentation in the epicontinental sea, when it was periodically connected with or isolated from the neighboring Arctic Ocean and northern marginal seas of the Tethys.

The first stage of Paleogene marine sedimentation begun after a short gap at the Maastrichtian–Danian boundary. The transgression of the early Talitsa time developed under active influence of the Arctic basin, and peculiar biogenic sedimentation in the trans-Urals structural zone suggests that the latter represented at that time a kind of upwelling zone in paleogeographic terms. This is evident from the fact that the Marsyaty

Formation thickness decreases from the north to the south within the area of the Ivdel' manganese deposit, and also from general distribution patterns of siliceous facies (opoka-like clays) of the lower Talitsa Subformation and of clayey to clay-calcareous Shadrinsk beds. The former are mainly localized in northern areas of the plate, whereas the latter are characteristic of its southern areas. In contrast, the late Talitsa transgression advanced from the south and this produced the corresponding impact on the benthic faunal assemblages. Larger calcareous and agglutinating foraminifers replaced smaller arenaceous foraminifers, which are characterized as primitive and dwarfed in description of Subbotina (*Foraminifery...*, 1964; Podobina, 1998). Silica accumulation ceased during this transgression, and northern plate areas accumulated coastal-marine and continental facies, i.e., sandy to clayey, frequently carbonaceous deposits of the upper Tibeisale Subformation. Assemblages of organic-walled phytoplankton and foraminifers (especially of calcareous forms) show that the Talitsa basin was tightly connected with sea basins of northwestern Europe during the Selandian time, as many species of these groups were originally described by Brotzen from the Selandian type sections of southern Sweden.

The second and longest stage of marine sedimentation commenced in the early Thanetian, and the regenerated silica accumulation was more intense than in the early Talitsa time. The onset of the Thanetian transgression improved connections between the Arctic and epicontinental sea basins, retaining communications of the latter with northern marginal seas of the Tethys. Intensification of meridional marine communications gave rise to the more intense heat and humidity transfer from lower to higher latitudes that resulted in climatic warming. This is evident not only from active bauxite formation in the Siberian platform at the Paleocene–Eocene boundary time and during the early Eocene, but also from concurrent development of broad-leaved thermophilic flora of the Tastakh type in the circum-Arctic areas of Siberia. During this short-term period, the entire northern hemisphere represented a kingdom of frost-free or alike climate. During the Irbit time of very intense biogenic silica accumulation, the West Siberian sea widened connections with the Arctic and Tethyan basins and entered into biogeographic communications with other seas of western Eurasia. In the Serov and Irbit periods of siliceous sedimentation, silica was supplied into the quasi-upwelling zone of the trans-Urals by underwater volcanic eruptions within the plate (Van, 1984) and by sea currents from the paleo-Arctic basin, where an intense underwater and subaerial volcanism was concentrated at the Paleocene–Eocene boundary time in its western areas (British-Arctic basaltic province). In the eastern zone of the plate, where fine terrigenous (clay) and siliceous components are mixed in sediments of the lower and middle Lyulin-Vor subformations, the sea current and silica-free waters were moving from the south to the north in contrast to the western plate margin, where the current streamed in opposite direction (Coriolis effect). The stream system of this kind easily explains the early Paleogene bauxite formation along the western margin of Siberian platform (Yenisei Ridge, Angara River basin, and Chadobets Uplift) and the extreme thickness of the Tavda Formation in the southeast of the study region. During the late Lyulin-Vor time, the Arctic basin gradually lost its effect on sedimentation in the epicontinental sea. Nevertheless, both sea areas kept connections for a long period, but the control of Tethyan waters over sedimentation and biota development was steadily growing. The West Siberian sea area was maximal in the terminal Ypresian–initial Lutetian time, when it was warm and occupies over 60% of the plate area (Shatskii, 1978). During the extensive regression of the Lyulin-Vor–Tavda boundary time, connections between Arctic basin and epicontinental sea ceased, areas of marine sedimentation considerably decreased, and older sediments experienced the subaerial to underwater erosion.

The Lutetian–Bartonian boundary marks the commencement of the third sedimentation cycle. The warm-water transgression of the early Tavda time developed from the south, as the synonymous formation is much thicker in the Kulunda–Barabinsk and circum-Kazakhstan structural zones than in other areas. As the Tavda Formation base is diachronous to some extent, the onset of marine sedimentation in the south slightly advanced that in the north. It is likely as well that siliceous sedimentation in southern areas ceased earlier, in the middle Eocene. Planktonic foraminifers characteristic of the Bartonian in the Crimea–Caucasus region and Central Asia migrated in West Siberia during the early Tavda transgression. During the entire second half of the Eocene, the sea under consideration was connected with southern sea basins. The second transgressive wave of the late Tavda (Priabonian) time left most distinct records in the south of the plate (Martynov, 1973). The same event is marked by compositional changes in the dinocyst assemblages from the upper Tavda Subformation penetrated by Borehole 011-BP in the Omsk district (Kul'kova, personal communication). After the Tavda sea desiccation at the Eocene–Oligocene boundary time, there was the short-term ingression of seawater from the south that resulted in deposition of the Kurgan Beds.

CONCLUSION

(1) The presented regional stratigraphic scheme of the West Siberian Paleogene has the following distinctions from other schemes formerly accepted or recently prepared for revision at the stratigraphic conference:

i. The pre-Paleogene Gan'kino Formation is omitted from the scheme. This unit is overlain with a hiatus by the Talitsa Formation, and basal beds of the latter correspond to the Danian.

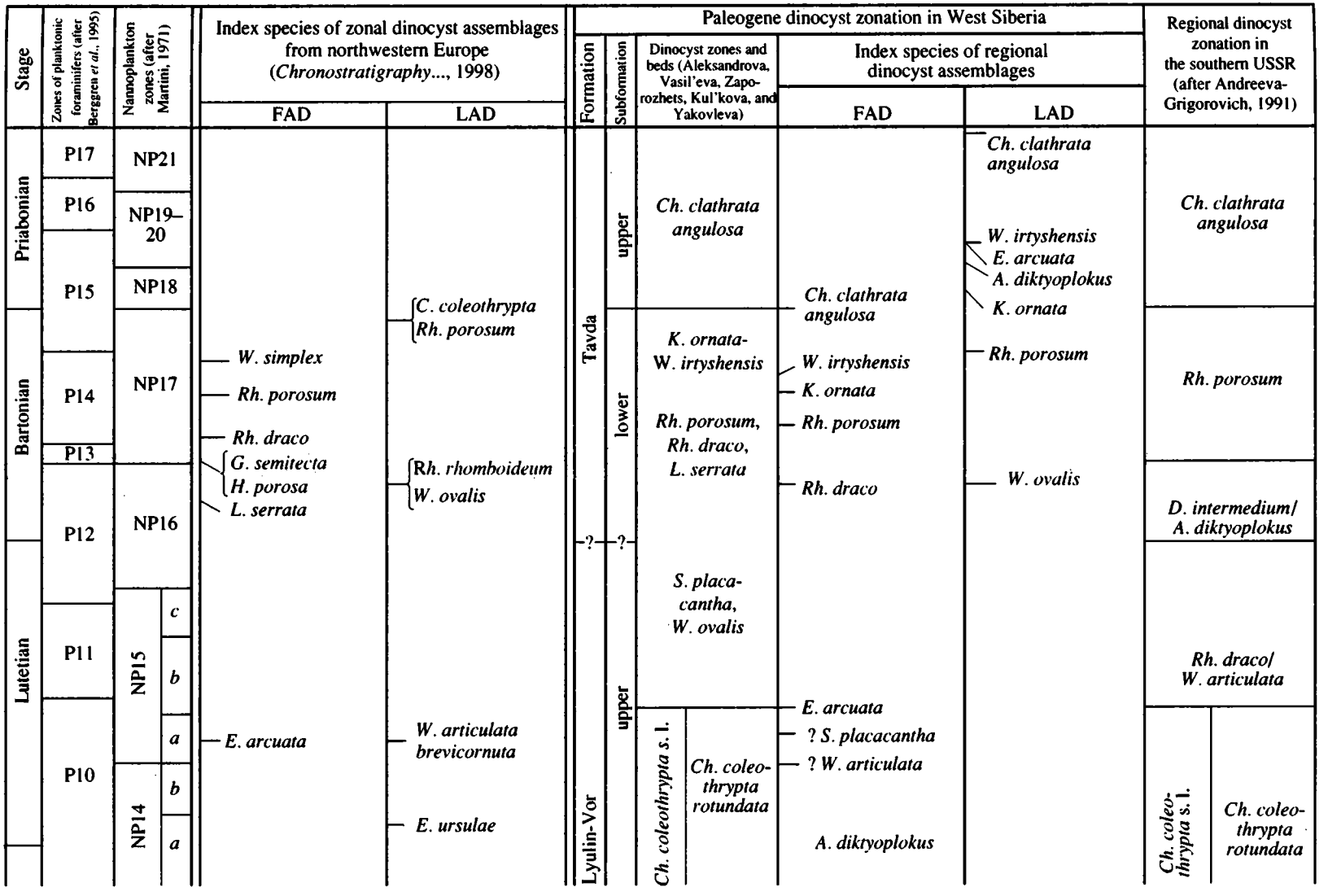


Fig. 4. Dinocyst zonations in West Siberia and southern USSR correlated with standard zones of planktonic foraminifers and nannoplankton (the first and last occurrence levels of dinocyst index species in the North Sea and West Siberia are shown in corresponding columns of the scheme).

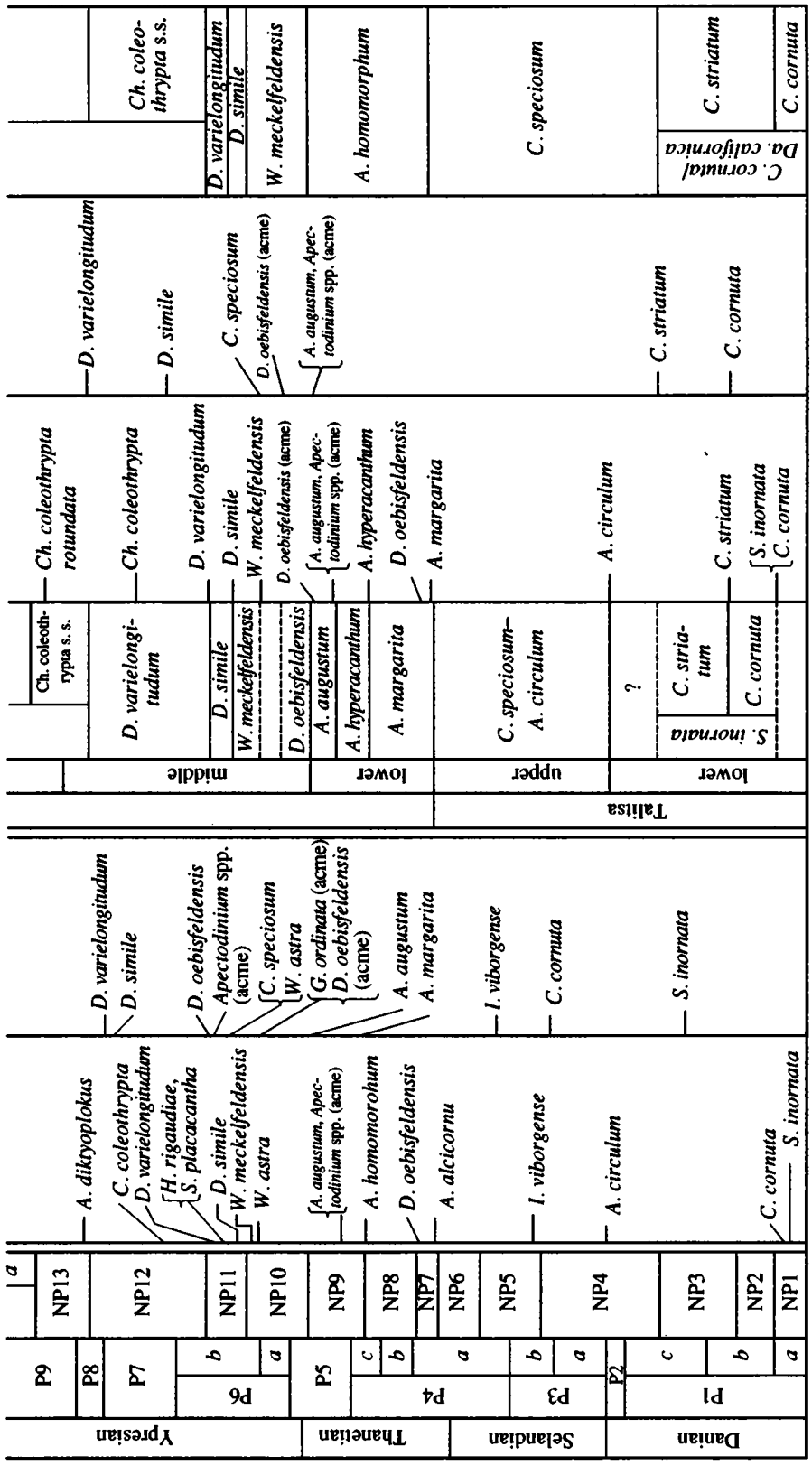

Ypresian
Thanetian
Selandian
Danian
P9
P8
P7
P6
P5
P4
P3
P2
P1
NP13
NP12
NP11
NP10
NP9
NP8
NP7
NP6
NP5
NP4
NP3
NP2
NP1
a
b
c
A. diktyoplokus
C. coleothrypta
D. varielongitudum
H. rigaudiae,
S. placacantha
D. simile
W. meckelfeldensis
W. astra
A. augustum, Apectodinium spp. (acme)
A. homomorphum
D. oebisfeldensis
A. alcicornu
I. viborgense
A. circulum
C. cornuta
S. inornata
D. varielongitudum
D. simile
D. oebisfeldensis Apectodinium spp. (acme)
C. speciosum
W. astra
G. ordinata (acme)
D. oebisfeldensis (acme)
A. augustum
A. margarita
I. viborgense
C. cornuta
S. inornata
Talisa
lower
middle
upper
lower
Ch. coleothrypta s. s.
D. varielongitudum
D. simile
W. meckelfeldensis
D. oebisfeldensis
A. augustum
A. hyperacanthum
A. margarita
C. speciosum–A. circulum
?
S. inornata
C. striatum
C. cornuta
Ch. coleothrypta rotundata
Ch. coleothrypta
D. varielongitudum
D. simile
W. meckelfeldensis
D. oebisfeldensis (acme)
A. augustum, Apectodinium spp. (acme)
A. hyperacanthum
D. oebisfeldensis
A. margarita
A. circulum
C. striatum
S. inornata
C. cornuta
D. varielongitudum
D. simile
C. speciosum
D. oebisfeldensis (acme)
A. augustum, Apectodinium spp. (acme)
C. striatum
C. cornuta
Ch. coleothrypta s.s.
D. varielongitudum
D. simile
W. meckelfeldensis
A. homomorphum
C. speciosum
C. cornuta/ Da. californica
C. striatum
C. cornuta

Fig. 4. (Contd.)

ii. The accumulation period of the Talitsa Formation spans the Danian and Selandian ages, and only the uppermost beds of the unit may be dated back to the Thanetian. Relationships between the lower and upper Talitsa subformations are insufficiently clear.

iii. Assemblages of organic-walled and siliceous plankton allowed us to establish quite confidently the late Eocene age of the lower Lyulin-Vor Subhorizon and the Ypresian–early Lutetian (?) time range of the middle and upper Lyulin-Vor subhorizons.

iv. The Tavda Formation and its age analogues accumulated in marginal zones of the plate during the Bartonian and Priabonian. Nevertheless, it is necessary to specify age of the formation base in those structural-facies zones, where rock sequences are lacking hiatus at the boundary between the Lyulin-Vor and Tavda formations.

(2) Our version of regional stratigraphic scheme suggested for the marine Paleogene in West Siberia is substantiated by the refined age data inferable for principal lithostratigraphic subdivisions from the correlated zonations of various microplankton groups. Correlating regional dinocyst assemblages with those known from the Danian, Selandian, Thanetian, and Ypresian stratotypes of the North Sea basin, we distinguished in the reference sections of West Siberian plate all the dinocyst zones of the Paleocene and lower Eocene, which characterize the West European standard and the respective zonal scales accepted for the Crimea–Caucasus region and East European platform. The established regional dinocyst zonation was directly correlated with standard scales based on planktonic foraminifers and nannoplankton (Fig. 4) that was used to refine time ranges of other faunal and floral groups and assemblages from West Siberia. Within the Lutetian interval, correlation between the regional and West European dinocyst zones is less obvious owing to the higher degree of endemism. Index species in common are known for the Bartonian and Priabonian zonal assemblages, but the West Siberian biota of the Tavda time was also more endemic than that of the Early Paleogene.

(3) The Talitsa, Lyulin-Vor, and Tavda horizons represent three regional subdivisions well traceable over the West Siberian plate. They should be considered as equal in rank within the hierarchy of regional stratigraphic units.

(4) The studied biostratigraphy and detailed stratigraphic scheme of marine Paleogene in West Siberia based on all fossil groups, except for the calcareous foraminifers, show that regional sections can be used as a basis for elaboration of the standard Paleogene scale for the Boreal region. In this case, dinocysts and siliceous microplankton represent fossil groups of prime significance.

(5) Among all unsolved problems of Paleogene biostratigraphy in West Siberia, the urgent one concerns the competent discrimination of the Tavda Formation from underlying deposits.

(6) The epicontinental sea basin of West Siberia was largest one among Paleogene seas of the Earth (area over 2 millions km^2). For a long period, this basin and Turgai strait connected cold- and warm-water zones of northern hemisphere acting as climatic regulator that ensured the heat and humidity transfer from lower to high latitudes. In this respect, the basin is a very appropriate object that can be used for evaluation of climate-forming significance of meridional seaways in the geological history of the Earth. In addition, this object offers a unique opportunity to study in detail the Early Paleogene quasi-upwelling zone up to 1000 km long, where immense volumes of opal–cristobalite rocks have been deposited.

(7) Reconstructing geological history, paleogeography, and climatic environments in West Siberia, we simultaneously decipher the Paleogene history of the Arctic basin. Marine Paleogene sequences and fossil assemblages of the West Siberian type are now discovered in the Pechora basin and Kara Sea. The effect of Paleogene connections between the West Siberian and Amerasian basins is obvious, but it is equally important to understand the concurrent influence of the Euramerican basin on sedimentation, biota, and climate in the region under consideration. This purpose appears to be realizable despite the fragmented state of the Paleogene sedimentary cover that is eroded in many areas of the Barents Sea bottom. With this approach, we can get a thorough insight into the mutual history of the West Siberian and Arctic basins.

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The Paleogene Reference Section of the Kril'on Peninsula (the Kitosiya River, Southern Sakhalin)

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Abstract—The article presents results of investigations of the Paleogene section at the Kitosiya River (the Kril'on Peninsula, southern Sakhalin). The section is proposed to be the reference one for the Kril'on subzone of the Southern structural–facies zone of the West Sakhalin depression originated in the Paleogene time. Lithological–facies peculiarities of the section are characterized, and lists of identified foraminifers, mollusks, plant imprints, spores and pollen are presented. The Kitosiya River section exhibits the lower Paleocene, middle–upper Eocene and Oligocene deposits. Stratigraphic units of the section are correlative to those of the Ishikari and Kushiro coal basins of the Honshu Island of Japan. Some formations are dated more accurately. After investigation of foraminifers, analogues of the Sinegorka Beds of the Dolinsk area are recognized for the first time at the base of the Paleogene section. The term "Snezhinka Formation" is shown to be preferable for the continental coal-bearing deposits of the Southern structural–facies zone. A stratigraphic hiatus is suspected at the Paleocene–Eocene boundary level.

Key words: southern Sakhalin, Paleogene, reference section, foraminifers, flora, mollusks, correlation.

INTRODUCTION

The Kril'on Peninsula (the southern extremity of western Sakhalin) is referred to the Kril'on subzone of the Southern structural–facies zone of the West Sakhalin depression formed in the Paleogene time (Khvedchuk *et al.*, 1970). L.S. Margulis and V.O. Savitskii who originally distinguished the subzone did not publish their results.

The Paleogene deposits are most completely exposed at the Kitosiya River[1] in the western part of the peninsula (Fig. 1). The lower part of the Paleogene section is predominantly composed of continental coal-bearing deposits with marine interbeds containing benthic foraminifers. The thicker upper part mainly consists of coastal-marine and deeper terrigenous–volcanogenic deposits. The weakly disturbed sequence of monoclinal beds shows distinct boundaries of the Paleogene deposits with underlying Cretaceous and overlying Miocene deposits and contains abundant paleontological remains (foraminifers, mollusks, plant imprints, spores, and pollen), which are well studied. The completeness and a high investigation degree of the Kitosiya River section allow us to consider it as the Paleogene reference section for the Kril'on subzone of the Southern structural–facies zone.

In addition, the close similarity between this and Paleogene sequences of Hokkaido in the depositional environments and composition of fossil assemblages is favorable for the detailed correlation of Paleogene deposits in these two neighboring regions and for more accurate dating of some formations of the Kitosiya River section.

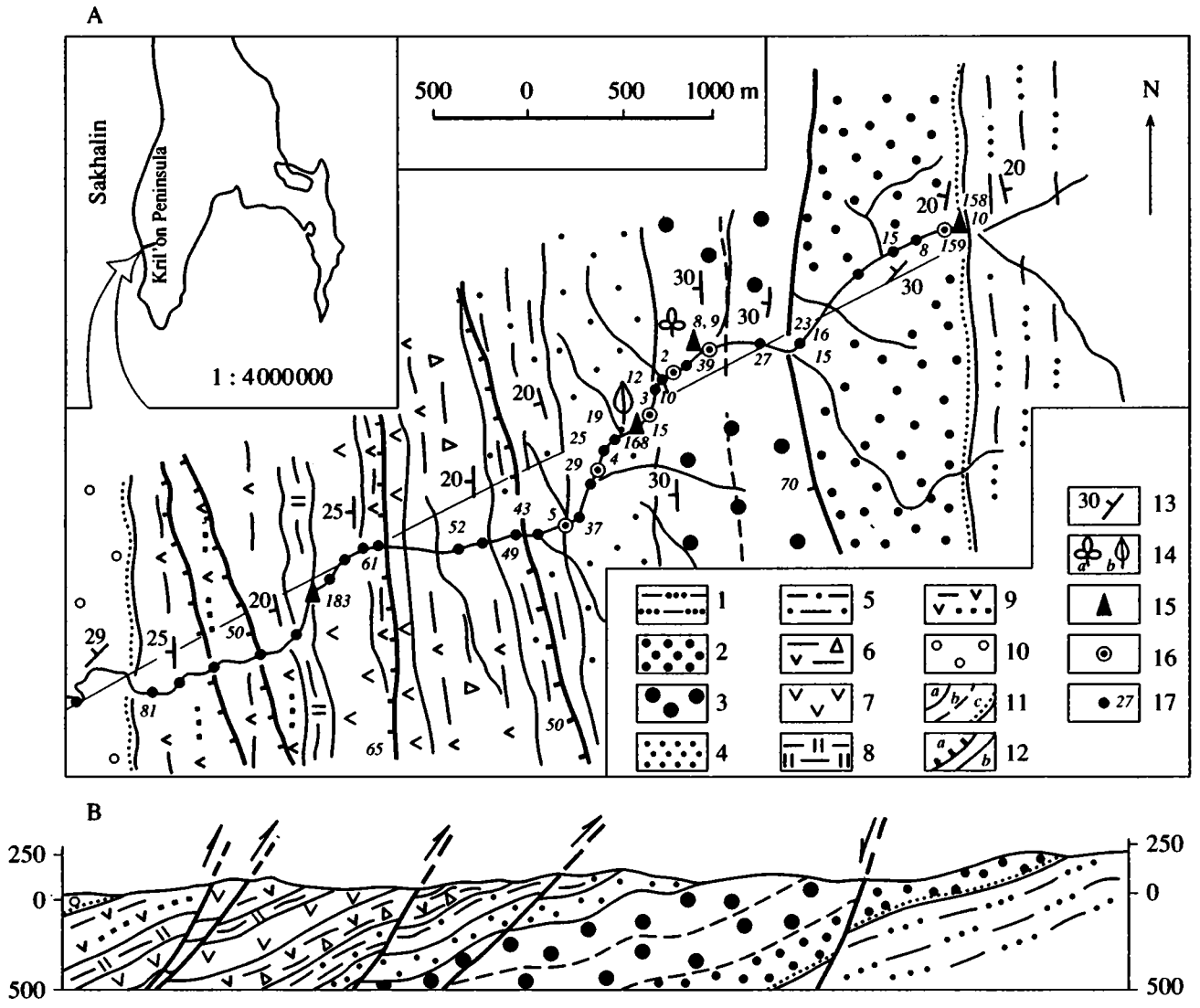
INVESTIGATION HISTORY OF THE KITOSIYA REFERENCE SECTION

The Japanese geologist K. Uwatoko was first to elaborate a unified stratigraphic scheme for Cenozoic deposits of southern Sakhalin, the Kril'on Peninsula area included (Smekhov, 1944). The scheme incorporated four groups subdivided into ten formations.

After the war time, an intense mapping and special investigations of southern Sakhalin including the Kitosiya reference section were carried out in 1969 by L.S. Margulis, V.O. Savitskii, N.M. Grokhotova, and other geologists from the Sakhalin Geological Survey (Yuzhno-Sakhalinsk), in 1980 by B.A. Sal'nikov, N.Ya. Brutman, V.N. Kuznetsova, L.A. Popova, and other researchers from the Sakhalin Department of the All-Union Petroleum and Geological Exploration Institute (Okha), and by geologists and paleontologists of

† Deceased.

[1] The river is also known under name the Chkalov (Il'ev *et al.*, 1977) or Chkalovka River (Medyulyanov, 1974; *Geologicheskie...*, 1996).


A
Sakhalin
Kril'on Peninsula
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1 Krasnoyarka Formation, Upper Cretaceous
2 Sinegorka Beds, lower Paleocene
3 Snezhinka Formation, lower middle Eocene
4 Krasnopol'e Formation, middle – lower upper Eocene
5 Takaradai Formation, upper Eocene
6 lower Arakai Subformation, upper Eocene
7 upper Arakai Subformation, Oligocene
8 Kholmsk Formation, lower Miocene
9 Nevel'sk Formation, lower Miocene
10 Verkhnii Due Formation, lower Miocene–lower middle Miocene
11 geological boundaries conformable between (a) and inside stratigraphic units (b), unconformable (c)
12 reversed thrust (a) and normal (b) faults
13 attitude of bedding
14 sites with early–middle Eocene (a) and late–middle Eocene (b) plant fossils
15 observation points of Savitskii and Margulis
16 observation sites of Fot'yanova
17 sampling sites by Fot'yanova and their numbers

Fig. 1. Location of the studied section in the geological map of the Kitosiya River basin: (1) Krasnoyarka Formation, Upper Cretaceous; (2) Sinegorka Beds, lower Paleocene; (3) Snezhinka Formation, lower middle Eocene; (4) Krasnopol'e Formation, middle – lower upper Eocene; (5) Takaradai Formation, upper Eocene; (6) lower Arakai Subformation, upper Eocene; (7) upper Arakai Subformation, Oligocene; (8) Kholmsk Formation, lower Miocene; (9) Nevel'sk Formation, lower Miocene; (10) Verkhnii Due Formation, lower Miocene–lower middle Miocene; (11) geological boundaries conformable between (*a*) and inside stratigraphic units (*b*), unconformable (*c*); (12) reversed thrust (*a*) and normal (*b*) faults; (13) attitude of bedding; (14) sites with early–middle Eocene (*a*) and late–middle Eocene (*b*) plant fossils; (15) observation points of Savitskii and Margulis; (16) observation sites of Fot'yanova; (17) sampling sites by Fot'yanova and their numbers.

the Sakhalin Complex Research Institute (Novo-Aleksandrovsk) (Zhidkova *et al.*, 1974; Il'ev *et al.*, 1977).

Beginning from 1990, V.G. Gal'versen, A.E. Zharov, and others from the Southern Sakhalin Mapping Team of the Sakhalin Geological Exploration Expedition studied the Kitosiya River area to prepare sheets of the State Geological Map-200. According to the agreement between the Sakhalin Geological Exploration Expedition and the Paleontological Institute of Russian Academy of Sciences, Fot'yanova performed in 1991 the stratigraphic and paleontological investigations of the Kitosiya section in order to date more accurately the continental coal-bearing deposits and overlying transgressive marine sediments.

The detailed study of biofacies and the bed-by-bed description of the Kitosiya River section were first made in the late 1960s by Margulis, Savitskii and other specialists from the Sakhalin Geological Survey. Regretfully, their materials have not been published. These specialists recognized above the Cretaceous deposits the continental Naiba Formation consisting of three (lower, middle and upper) subformations and showing lithological similarity with the synonymous formation of the Dolinsk stratotype area of southern

Sakhalin (Smekhov, 1947). They did not distinguish the Krasnopol'e Formation represented, in their opinion, by similar facies in this section. The Naiba Formation is conformably overlain by the Aral'skaya Formation crowned with the Nikiforovka Beds. The succeeding and conformable Kholmsk Formation begins the Neogene sequence of the Kitosiya River section. Later all researchers followed this subdivision scheme though correcting scopes and ages of the formations and using different nomenclature of the units, especially of those recognized in basal continental deposits. For instance, Zhidkova *et al.* (1974), Il'ev *et al.* (1977), and other specialists attributed these deposits to the Nizhnii Due Formation. Kodrul (1994) distinguished the Paleogene continental deposits of the Kril'on Peninsula, including those of the Kitosiya section, as the Naiba Formation. When studying the Cretaceous/Paleogene boundary of Sakhalin, Vereshchagin *et al.* (1978) suspected the occurrence of the Sinegorka Beds analogues under the Nizhnii Due Formation. However, the supposition was not supported by paleontological evidence.

In the upper part of the Nizhnii Due Formation of the Kitosiya River section, Zhidkova *et al.* (1974) and Il'ev *et al.* (1977) recognized on the basis of mollusks the Krasnopol'e Formation of an insignificant (60 m) thickness. According to the general opinion, the Takaradai Formation conformably overlies the Krasnopol'e Formation, but Savitskii and Margulis considered the latter and the Arakai Formation, as well as the Nikiforovka Beds, as stratigraphic equivalents of the Takaradai Formation that is conformably overlain by the Arakai Formation divided into two subformations. The last formation underlies the Kholmsk Formation, the lower unit of the Neogene sequence.

Summarizing the brief historical review, we should note that the main problem in subdividing the Kitosiya River section concerns nomenclature and scopes of its Paleogene formations.

This work is based on the material collected by Fot'yanova in 1991, when she visited the Kitosiya section that was characterized by Sal'nikov (Sakhalin Marine Petroleum Scientific and Production Institute, Okha) as very well exposed and relatively accessible in the western Kril'on Peninsula. The foraminiferal and palynological analyses were performed using the same rock samples. Paleontological determinations were made by Serova (foraminifers), Tuzov (mollusks), Fot'yanova (flora), and Grokhotova (spores and pollen). The molluscan assemblages were distinguished by Tuzov on the basis of samples collected by Savitskii and Margulis from the Kitosiya River section and also from some data published earlier (Zhidkova *et al.*, 1974; Zakharova and Sheremet'eva, 1977). The results obtained are presented below.

DESCRIPTION OF THE PALEOGENE REFERENCE SECTION, THE KITOSIYA RIVER SITE

The following description is given according to Margulis and Savitskii with additional data of Gal'versen, Zharov, and Fot'yanova.

The section exhibits the following units (from the base upward):

The Krasnoyarka Formation. The upper part of the formation near the contact with the Sinegorka Beds is composed of glauconite-bearing coarse- to medium-grained sandstones.

The Sinegorka Beds. The beds rest on the erosional surface of the Upper Cretaceous Krasnoyarka Formation. In the Lugovka River basin, N.P. Voronin (the Sakhalin Geological Exploration Expedition, Yuzhno-Sakhalinsk) distinguished an unconformity between the Sinegorka Beds, which he referred to the Nizhnii Due Formation, and the Upper Cretaceous deposits. The beds include the basal conglomerate interbed about 10 m thick. The Sinegorka Beds of the Kitosiya section comprise three members (1–3 in Fig. 2, from the base upward).

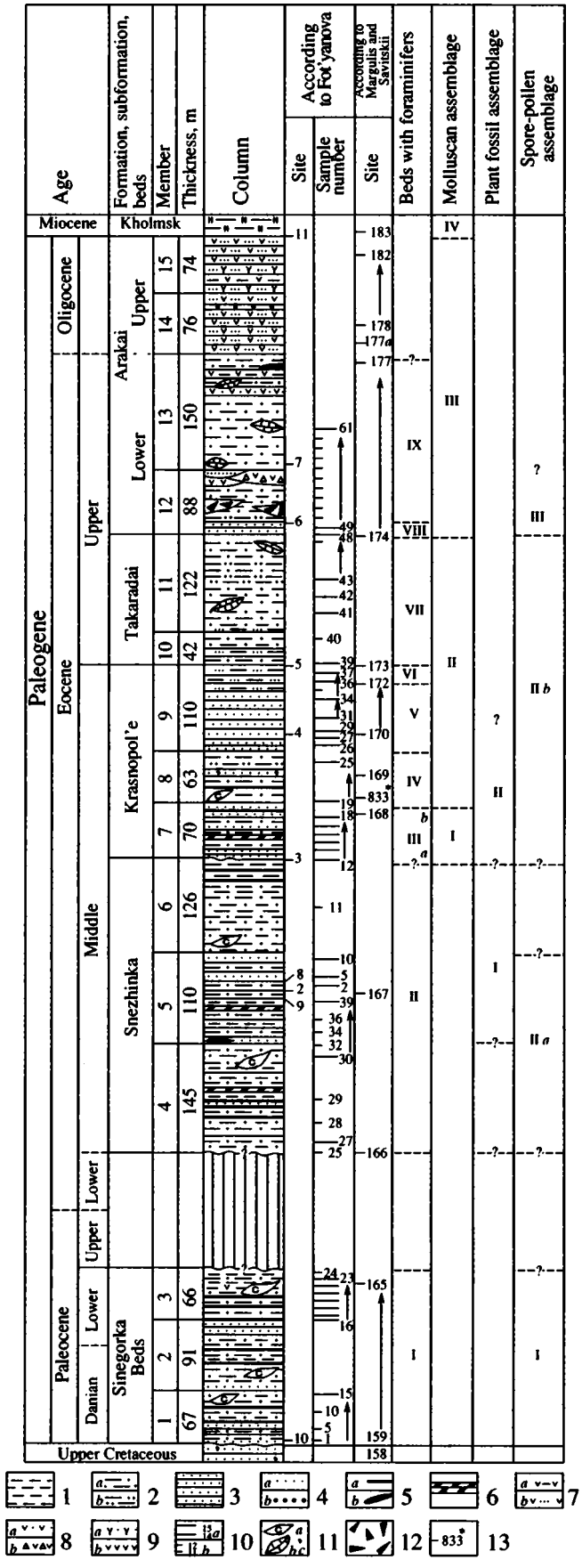
Member 1 (slightly coaliferous; Site *10*, samples 1–15).[2] The member is mainly composed of dark gray massive and uniform siltstones with three coal seams 0.15, 0.5, and 1.1 m thick. The siltstones contain thin interbeds of greenish gray medium- to fine-grained sandstones and siderite concretions. A bed (about 3 m thick) of gravelstone and coarse-grained sandstone occurs at the base of the member. The dark gray siltstones at the top (Site *10*, Sample 15) yielded foraminifers *Haplophragmoides granulosum* Lipman, *H. retrosepta* (Grzyb.), *Ammobaculites expansus* Plum., and *Silicosigmoilina longa* Serova. Impressions of fern *Osmunda sachalinensis* Krysht. are detected at the levels of 30.0 and 51.0 m above the base of the member.[3] The member is 67.0 m thick.

Member 2. Two beds of light gray fine-grained sandstones about 11.0 (lower) and 13.0 m (upper) thick are separated by siltstone sequence (about 67.0 m) with mudstone interbeds, two coal seams, and relatively scarce concretions of siderite sandstones. The member is 91.0 m thick.

Member 3. Dark gray siltstones containing mudstone interbeds and siderite concretions at the top and enclosing three coal seams; the member (Samples 16–23)[3] contains agglutinating foraminifers: abundant *Haplophragmoides retrosepta* (Grzyb.), *H. granulosum* Lipman, *Haplophragmoides* sp. and single *H. eggeri* Cushm. and *Asanospira glabra* (Cushm. et Waters).

[2] Site and sample numbers are given according to Fot'yanova.

[3] Fossil plants are identified by Medyulyanov (the Sakhalin Geological Exploration Expedition, Yuzhno-Sakhalinsk).



1 2 3 4 5 6 7 8 9 10 11 12 13

Fig. 2. The Paleogene reference section at the Kitosiya River: (1) mudstone; (2) siltstone (*a*) or sandy siltstone (*b*); (3) sandstone; (4) gravelstone (*a*) or conglomerate (*b*); (5) coal seam (*a*) or lens (*b*); (6) carbonaceous mudstone; (7) tuffaceous siltstone (*a*) or sandstone (*b*); (8) psammitic (*a*) or psephitic (*b*) tuff-agglomerate of intermediate composition; (9) tuffite (*a*) or intermediate and basic lavas (*b*); (10) number of samples containing (*a*) or devoid (*b*) of foraminifers; (11) siderite (*a*), marlaceous (*b*), or glauconite-bearing (*c*) concretions; (12) mixtite; (13) sampling site of Gal'versen.

Members 1–3 are characterized by Samples 1–24 (Site *10*) and by observation points 159–165.[4] The Member 3 is 66.0 m thick.

In total, the Sinegorka Beds are 224.0 m thick.

The coaliferous Snezhinka Formation.[5] Margulis and Savitskii observed an azimuthal discordance between the lower part of the Naiba Formation (points *159–165*; the Sinegorka Beds in our interpretation) and its middle and upper parts (points *166* and *167*), although both the Sinegorka Beds and the Naiba Formation exhibit the same dipping angles of about 20–30°. Gal'versen and Zharov distinguished closely spaced submeridional faults at the contact of the Sinegorka Beds with the Snezhinka Formation (Point *166*; Fig. 1). They supposed that a part (not less than 100 m) of the Snezhinka Formation is missing from the Kitosiya River section.

The fault amplitude was not determined with confidence. It was graphically reconstructed and consistent with the general thickness estimation (800–850 m) of the continental deposits in the Lugovka and Shebuninka river basins.

The formation is divided into three members (4–6 from the base upward):

Member 4. This coal-bearing member is mainly composed of gray coarse-grained siltstone locally grading into sandstone and intercalated with interbeds of mudstone and carbonaceous mudstones; the maximal thickness of lenses and five coal seams is 3.0 m. The member contains abundant siderite concretions. In the middle part of the member, the latter reveal single imprints of *Corylus* sp., *Platanus* sp., and unidentifiable fragments of dicots. Foraminifers occur sporadically, and only two samples 27 and 30 at Site *10* yielded abundant but deformed and poorly preserved thin-walled *Trochammina* tests conventionally referred to *Trochammina vitrea* Serova, single fragments of *Rzehakina* sp. and *Glomospira gordialis* Cushm. et Jarv. in association with sponge spicules. The apparent thickness of the member is 145.0 m.

Member 5. This slightly coaliferous member begins with the thick (up to 14.0 m) bed of light gray

[4] The points of Margulis and Savitskii.

[5] The correspondence of coal-bearing deposits to the Snezhinka Formation is substantiated below in the section "Nomenclature of continental coal-bearing deposits of the West Sakhalin Mountains."

medium-grained sandstone enclosing one coal lens up to 0.4 m thick. Analogous sandstone bed (up to 38.0 m thick) with thin interbeds of gray siltstone crowns the member. The sandstone beds are separated by a sequence of alternating dark gray mudstones, light gray fine-grained sandstones, less frequent siltstones, and carbonaceous mudstones. These deposits include four thin coal seams. Only two siltstone samples yielded abundant though deformed tests of *Trochammina* sp. and sponge spicules. In the upper part of the member, mudstones bear plentiful unidentifiable pyritized casts of *Haplophragmoides* sp., single deformed thin-walled tests of *Trochammina* sp., and specimens of *Silicosigmoilina futabaensis* Asano (Site *10*, Sample 39; Site 2, samples 2, 5, and 10). The fossil plant assemblage consisting of *Metasequoia occidentalis* (Newb.) Chaney, *Trochodendroides speciosa* (Ward) Berry, *Platanus* sp. (with peltate leaf base), *Quercus* sp., *Fagus* sp., *Castanea? longifolia* Borsuk, *Alnus* sp., and *"Cordia" japonica* Tanai[6] was found in siltstones underlying the upper sandstone bed of Member 5 (Site *9* and Point *167*). *Metasequoia occidentalis* (Newb.) Chaney, *Trochodendroides speciosa* (Ward) Berry, *Magnolia kryshtofovichii* Borsuk, *Platanus* sp., *Populus* cf. *celastrophylla* (Bail.) Sycheva, *Tilia* cf. *setadentata* Sycheva, *Fagus* sp., *Alangium basitruncatum* (Oishi et Huzioka) Tanai, and *"Cordia" japonica* Tanai were discovered at Site *8* approximately 20 m above the Site *9*. The member is 110.0 m thick.

Member 6 (slightly coaliferous). The lower part of the member (about 50 m) corresponds to gray uniform siltstones with siderite concretions. Higher in the section, siltstones include mudstone interbeds bearing plant remains of *Metasequoia disticha* (Heer) Miki, *Ulmus furcinervis* (Borsuk) Ablaev, and *Trochodendroides arctica* (Ung.) Heer[3]. A thick bed of homogenous gray mudstone with rare thin lenses and two coal seams is topmost in the member. The thickness of Member 6 is 126.0 m. The apparent thickness of the Snezhinka Formation (Members 4, 5, 6) is up to 380.0 m.

The Krasnopol'e Formation has a conformable or erosional contact with the Snezhinka Formation. Lithologically the formation can be subdivided into three members (7–9 from the base upward).

Member 7. The unit is represented by alternating light gray fine-grained sandstones and siltstones. The middle part of the member includes a carbonaceous mudstone interbed 0.3 m thick. The section is crowned by a bed of fine-grained sandstone with a coal seam 0.5 m thick. The member (Site *3*, Samples 12–18 and Point *168*[4]) contains the most diverse foraminiferal assemblage dominated by arenaceous forms. The most abundant forms (more than 100 specimens in some samples) are *Haplophragmoides tanai* Kaiho and *H. subevolutus* Kaiho. The assemblage also includes plentiful *Haplophragmoides coalingensis* Cushm., less frequent *Rhabdammina eocenica* Cushm. et Hanna and *Reophax tappuensis* Asano, and single *Trochammina asagaiensis* Asano and *Trochammina* sp. (deformed specimens). More diverse and well-preserved calcareous forms are represented by single specimens. *Quinqueloculina reicheli* Le Calvez, *Q. yeguaensis* Weinz. et Appl., *Fissurina marginata* Montagu, *Cassidulina yubariensis* Kaiho, *Globocassidulina globosa* (Hantk.), *Nonion* aff. *durchami* Mall., *Elphidium ishikariense* Kaiho, *E. asanoi* Kaiho, and *Rosalina douvilleri* (Cushm.) were found in Sample 12, Site *3*[2]. The member is 70.0 m thick.

Member 8. Siltstones and gray sandy siltstones with an interbed of greenish gray glauconitic sandstones and some coal interbeds up to 0.3 m thick are constituents of the member. Siderite concretions occur in abundance. The lower part includes laminae rich in gravel and small pebble. Being still frequent (Site *3*, samples 19–25)[2], foraminifers reveal the sharply reduced abundance and taxonomic diversity. They are represented by one secreting species *Nonion takayanagii* Kaiho and by scarce agglutinating forms *Haplophragmoides takayanagii* Kaiho, *H. subevolutus* Kaiho, and *Haplophragmoides* sp., which have coarser grained walls than the same forms from underlying beds. A pelecypod assemblage consisting of *Nuculana (B.) pulchrisinuosa* Clare, *Nuculana (N.)* sp., *Yoldia takaradaensis* L. Krysht., *Yoldia* sp., and *Corbicula* sp. was found approximately 40 m above the member base. Gal'versen detected a leaf imprint of *Macclintockia pugetensis* Wolfe in the basal sandstones (Site *833* of Gal'versen, Fig. 2). The thickness of the member is 63.0 m.

Member 9. The unit consists of dark gray fine-grained sandstones, which are clayey in the upper part of the member. Sporadic foraminifers are represented only by agglutinating forms (Site *3*, samples 26, 27)[2]. A characteristic feature is the occurrence of *Cyclammina pacifica* Beck. and *C. samanica* Berry unknown from the lower part of the section. The samples also contain *Rhabdammina eocenica* Cushm. et Hanna, *Ammodiscus parianus* Hedberg, *Haplophragmoides coalingensis* Cushm. et Hanna, *Haplophragmoides* aff. *kirki* Wick., *Ammobaculites akabiraensis* Asano, *"Ammobaculites* sp." Mallory, and abundant *Haplophragmoides crassaformis* Kaiho. Higher in the section (Site *4*, samples 29–31, 33, 34, 37)[2], there are *Reophax minutiformis* Kaiho, *R. multicameratus* Kaiho, *Haplophragmoides takayanagii* Kaiho, *Budashevaella curviseptata* (Budash.), and other forms. Point *170*[4] yielded mollusks *Nuculana (N.)* sp., *Nuculana (B.) polahrisinuosa* Clare, *Yoldia takaradaensis* L. Krysht., *Venericardia petvensis* Laut., *Mya* sp., *Nemocardium weavery* (And. et Mart.). Further upward (Point *171*)[4], there were identified *Venericardia petvensis* Laut., *Mya* sp., *Nemocardium weavery* (And. et Mart.), *Macoma* sp., and *Mya sachalinensis* L. Krysht. The only species *Nuculana (N.)* sp. was found at the top of the member (Point *172*)[4]. The member is 110.0 m thick.

[6] Fossil plant identifications by Fot'yanova.

The total thickness of the Krasnopol'e Formation (Sites *168–172*)[4] is 243.0 m.

The Takaradai Formation. The formation consists of two members (10 and 11 from the base upward).

Member 10 is represented by gray fine-grained clayey sandstones in the lower part and by dark gray homogenous, locally sandy siltstones in the upper part. The basal beds (Site *5*, Sample 39)[2] yielded abundant and deformed *Trochamina* sp. and single specimen of *Nonion takayanagii* Kaiho. The member contains mollusks *Nemocardium weavery* (And. et Mart.) and *Liocyma* sp. The thickness of the member is 42.0 m.

Member 11 uniformly consists of bluish to dark gray, locally sandy siltstones. The relatively diverse foraminiferal assemblage from the middle part of the member contains agglutinating and predominant secreting forms. Among them, there are abundant *Haplophragmoides yokoyamai* Kaiho and *H. aperturiumbilicatus* Kaiho, less frequent *Reophax tappuensis* Asano and *Trochammina* sp., single *Cyclammina pacifica* Back., *Budashevaella curviseptata* (Budash.), *Nonion takayanagii* Kaiho, and *Fissurina* sp. (Site *5*, samples 41–43)[2]. Mollusks *Nemocardium weavery* (And. et Mart.), *Mya sachalinensis* L. Krysht., *Acila (Truncacila)* sp., *Megayoldia* cf. *nitida* Slod., and *Periploma kariboensis* L. Krysht were found in the lower part of the member. *Nuculana (N.)* sp., *Acila (Truncacila)* sp., and *Periploma kariboensis* L. Krysht. were detected in the middle part, and *Nemocardium weavery* (And. et Mart.), *Ennucula* cf. *karatsuensis* (Khom.), *Acila* cf. *praesinnaensis* L. Krysht., and *Megayoldia cerussata* Slod. were identified in the upper part. The thickness of the member is 122.0 m. The total thickness of the Takaradai Formation is 164.0 m.

The Arakai Formation rests on the Takaradai Formation and consists of two, lower (members 12 and 13) and upper (members 14 and 15) subformations.

The lower subformation, Member 12. The member has a basal of gravelstone to coarse-grained sandstone horizon grading upward into siltstones, locally sandy ones. The siltstones include lenses, lenticular interbeds, and irregular bodies of gravitational mixtites 1.5–2.5 m, contacts with which are conformable or slightly discordant. Mixtites are coarse, poorly sorted, and enclosing differently rounded rock clasts (sand and gravel, pebbles, rubble, blocks, and boulders) set in silty and psammitic matrix. In contrast to the enclosing deposits, the mixtite matrix contains a significant volcanomictic and siliceous admixture. At the top of the member, there is a lens of tuff agglomerate and flow of basaltic andesite up to 8.0 m thick. The basal portion (Site *5*, Samples 48 and 49) identified by Savitskii and Margulis as the "Nikiforovka Beds" bear abundant *Budashevaella curviseptata* (Budash.), *Cyclammina ezoensis* Asano, and *Haplophragmoides evolutus* Kaiho, which dominate in the foraminiferal assemblage. *Cyclammina pacifica* Beck. is less frequent, while *Bathysiphon eocenicus* Cushm. et Hanna, *Ammodiscus tenuis* Bradv., and *Cyclammina incisa* Stache. occur sporadically. Mollusks *Megayoldia* cf. *nitida* Slod., *Nuculana (Borissia)* sp., *Malletia onorensis* Laut., *Laternula* cf. *besshoensis* (Yok.), *Acila (Truncacila)* cf. *osugii* (Nag. et Huz.), and *Venericardia* cf. *tokunagai* Yok. were found in the lower part of the member (Point *174*)[4]. The member is 88.0 m thick.

Member 13 is represented by bluish gray and gray relatively uniform siltstones with marlaceous concretions. There are glendonites in the lower part, scattered pebble of effusive deposits in the middle part, and interbeds of dark gray psammitic tuffs, tuffaceous siltstones with floating pebble and coal lenses in the upper part of the member.

Only one sample (Site *7*, Sample 61)[2] contains foraminifers. Their taxonomically diverse assemblage is dominated by secreting species. the assemblage includes *Haplophragmoides yokoymai* Kaiho, "*Discammina* sp." A. Kaiho, *Lagena striata* d'Orb., *Fissurina marginata* Walker et Boys, *Oolina simplex* Reuss, *Dentalina catenula* Reuss, *Stilostomela lepidula* Schwager., *Plectofrondicularia smithi* Kaiho, and other forms.

Two molluscan species *Acila (Truncacila)* sp. and *Malletia onorensis* Laut. were identified at the Point *177*. The member is 150.0 m thick. The thickness of the lower Arakai Subformation is 238.0 m.

The upper subformation. Member 14. This member of dark gray fine- to coarse-grained tuffaceous sandstones contains in its lower part the "floating" pebble and an interbed of boulder conglomerates up to 2 m thick in the upper one. *Periploma kariboensis* L. Krysht. was found at the Point *177a* and *Venericardia* cf. *tokunagai* Yok., *Thracia condoni* Dall., and *Turcicula sachalinensis* L. Krysht. was discovered at the Point *178*. The thickness of the member is 76.0 m.

Member 15. Its lower part is composed of psammitic tuffites with an interbed of bluish gray siltstone. The upper part consists of finely alternating tuffites, tuffaceous sandstones, tuffaceous siltstones, and tuffs. The member is 74.0 m thick. The thickness of the upper Arakai Subformation is 150.0 m, and the formation proper is 388.0 m thick in total.

The Kholmsk Formation. This unit conformably overlies the Arakai Formation and is composed of uniform compact gray siltstones, locally siliceous and enclosing thin interbeds of ashy tuffs. The formation yielded two molluscan species *Solemya tokunagai* Yok. and *Macoma* cf. *simizuensis* L. Krysht. The apparent thickness of the formation is 70.0 m.

NOMENCLATURE OF THE PALEOGENE CONTINENTAL COAL-BEARING DEPOSITS OF THE WEST SAKHALIN DEPRESSION

When describing the Kitosiya section, we referred the coal-bearing deposits with coastal-marine interbeds, which overlie the Sinegorka Beds, to the

Snezhinka Formation that was previously recognized only in the Uglegorsk area of Sakhalin. To justify their affinity with the Snezhinka, but not Nizhnii Due or Naiba formations, we should consider briefly the stratotypes of these formations and paleontological substantiation of their ages.

In the northern, central and southern (including the Kril'on subzone) structural–facies zones of the West Sakhalin depression (Khvedchuk *et al.*, 1970), the following lower Paleogene subdivisions of terrigenous coal-bearing deposits can be recognized: Kamenka (=Konglomeratnaya) and Nizhnii Due formations in the northern zone, Snezhinka Formation in the central zone, and Naiba Formation in the southern zone. Owing to the lithological similarity and presumably close ages, these three coal-bearing formations are frequently united into the single Nizhnii Due Formation (Zhidkova *et al.*, 1974; Il'ev *et al.*, 1977; and others). In the Kitosiya section, these deposits were attributed either to the Naiba (Medyulyanov, 1974; Kodrul, 1994) or Nizhnii Due formations equivalent in age and facies composition (Il'ev *et al.*, 1977). Their relationships and scopes are subjects of continuous discussions, because fossil floras of the northern and southern structural–facies zones have not been properly described.

In the northern zone, Borsuk (1956) identified eight species (among them two species of open nomenclature and three new forms) in the stratotypes of the Kamenka (Konglomeratnaya) and Nizhnii Due formations located to the south of the Ogorodnaya Ravine, although the fossil locations have not been specified. Following the Krishtofovich (1936), Borsuk assigned the Nizhnii Due Formation to the upper Eocene–lower Oligocene. In all later publications concerning flora and age of the Nizhnii Due Formation in the stratotype section (Medyulyanov, 1974; Krasilov and Kundyshev, 1982; Krasilov *et al.*, 1986), there were presented only lists of plant remains. Krasilov and Kundyshev (1982) concluded that the Nizhnii Due Formation is of the Oligocene age. Later these researchers (Krasilov *et al.*, 1986) placed the Eocene–Oligocene boundary inside the Nizhnii Due Formation, approximately between its coal-free and coal-bearing parts. They proposed for the regional stratotype of the Eocene–Oligocene boundary in continental deposits the section located to the south of the Ogorodnaya Ravine. Paleomagnetic records (Krasilov *et al.*, 1986) supported this boundary position. Paleofloral data, however, does not permit to locate unambiguously the Eocene–Oligocene boundary in this section. First, the Eocene and Oligocene plant assemblages are incommensurable in species number (5 and 35 respectively); second, four species occur both in the Eocene and Oligocene assemblages thus being transient; third, the Oligocene assemblage contains many Paleocene and Eocene species, such as *Platycarya hokkaidoana* Tanai, *Platanus aceroides* Goeppert, *Trochodendrocarpus arcticus* (Heer) Krysht., and *Aesculus magnificum* (Newb.), Iljinskaja (the dominant species of the Paleocene flora of Kamchatka), as well as *Trochodendroides arctica* (Heer) Berry and *Acer arcticum* Heer (dominants of the Paleocene–Eocene floras). In the stratotype, the Nizhnii Due Formation probably does not exceed in scope the Eocene, corresponding to its middle subseries. Sycheva (1977) correlated the Nizhnii Due Formation of the Aleksandrovsk area (the northern zone) with the upper Snezhinka Subformation of the lower Eocene (the central zone).

Savitskii *et al.* (1974) distinguished the Snezhinka Formation in the central structural–facies zone of the West Sakhalin depression (the Uglegorsk area). Its stratotype was chosen at the Snezhinka River site. The formation was distinguished on the basis of enclosed floral remains studied by Sycheva in the early 1970s (Sycheva, 1977). In her opinion, the Snezhinka floral assemblage indicates a longer age range of the formation (Danian–Eocene) than that of the Nizhnii Due Formation in the stratotype section. In the lower Snezhinka Subformation of the Snezhinka River section, Sycheva (1977) identified some Late Cretaceous genera passing into the Paleocene deposits and accordingly treated as transitional ones. These are genera and species of Platanaceae: *Credneria* sp. (Plate VII, Fig. 1, 1a); *Aspidiophyllum latifolium* (Holl.) Sycheva (Plate VIII, Fig. 1; Plate IX, Fig. 2); *Protophyllum savitskii* Sycheva (Plate V, Figs. 1–3). Owing to the association of these taxa with many others, predominantly Eocene species, the Snezhinka sediments seemingly accumulated during the long Danian–Eocene period.

In 1990, Fot'yanova studied the Sycheva's collection from the Snezhinka stratotype of the Snezhinka River section and collections from the Snezhinka sections at the Kama and Aral'skaya rivers.[7] As it turned out to be, the forms of genera *Credneria* and *Aspidiophyllum* represent in fact the genus *Platanus*, whereas the peltate-based species of the genus *Protophyllum* belong to the genus *Alnus*. Analyzing the published data, Chelebaeva and Bratseva (1985) arrived at the same conclusion on the systematic position of above taxa.

Analysis of floral assemblages from the lower, middle and upper Snezhinka subformations recognized in the Snezhinka River section shows that the early assemblage includes remains of plants, which grew predominantly near sedimentary basins (*Platanus, Populus, Hicoria, Trochodendroides*, etc.). In the middle assemblage, these plant remains are less abundant and associate with newly appeared Fagaceae (*Quercus rectinervis* Borsuk, *Q. ussuriensis* Krysht.). The late assemblage is characterized by the reduced abundance of Fagaceae and by sharply increased proportion of Betulaceae represented by seven species of genera *Alnus, Betula, Carpinus*, and *Corylus*. Thus, in the Snezhinka River section, we can distinguish only two, the early-middle (united) and late Snezhinka floral assemblages. The former shows no taxonomic similarity to the Paleocene

[7] The collections are stored under no. 2 at the Museum of the Far East Center of Geological Information, Yuzhno-Sakhalinsk.

flora of West Kamchatka, being very close to the earliest middle Eocene Harutori flora of the Kushiro coal basin of Hokkaido (Tanai, 1970; Okada and Kaiho, 1992; Kodrul, 1994). Therefore, this assemblage is supposed to be of the early–middle Eocene age. After some corrections, the comprehensively studied Snezhinka floral assemblage (Sycheva, 1977) can be considered as the reference flora of the basal middle Eocene deposits of Sakhalin.

Smekhov (1947) distinguished the Naiba Formation in the eponymous river basin. The unit is an equivalent of the Naibuchi Formation recognized in 1934 by the Japanese geologist Kawasaki in the same area. The formation rests on the upper Paleocene Sinegorka Beds (Kalishevich and Posyl'nyi, 1958; Kalishevich *et al.*, 1981; Serova, 1987) probably with a significant hiatus at the base. In the Krasnoyarka River basin, Medyulyanov (1974) distinguished four floral assemblages (lists of taxa without descriptions and illustrations, except for the genus *Platanus*, were published by Medyulyanov in 1975). He also mentioned that the same four floral assemblages characterize the Naiba Formation of the Chkalova River section, but lists of taxa have not been published. In conclusion, Medyulyanov (1974, p. 73) wrote: "We abstain from correlating the paleofloras of the southern zone to those of the central and northern zones. The floral assemblages of these zones need to be more carefully studied. We also abstain from their correlation to the standard stratigraphic units."

Thus, we had an opportunity to compare our collection from the Kitosiya River section with collections from the Snezhinka Formation of the stratotype and other sections of the central structural–facies zone and also with taxa of this formation, which are comprehensively described by Sycheva (1977). Taking into account the lithological similarity between the Snezhinka and Naiba formations in the Kitosiya River section (although the Naiba deposits accumulated in settings closer to marine ones), we attribute the continental coal-bearing deposits of the Kitosiya section to the Snezhinka Formation, as it was suggested by Gal'versen and Zharov. This can be, however, the temporary decision. It cannot be excluded that a deeper insight into flora from the Nizhnii Due Formation distinguished in the Ogorodnaya Ravine stratotype may reveal its similarity with floras of the Snezhinka and Naiba formations. In this case, a uniform nomenclature can be used for the Eocene continental coal-bearing deposits of the entire West Sakhalin depression.

PALEOGENE BIOTA FROM THE KITOSIYA RIVER SECTION

Foraminifers. Foraminifers occur sporadically in the Cenozoic deposits of the Kitosiya River section. They were found in 55 of 189 samples. The scarce and low-diverse assemblages are dominated by agglutinating forms, especially in the lower and upper parts of the section. More diverse secreting forms are usually represented by single specimens. As a rule, the assemblage ranges correspond to sedimentation cycles and stages in basin evolution (Serova, 1978). The section reveals nine beds with peculiar foraminiferal assemblages.

(I) The *Silicosigmoilina longa–Haplophragmoides granulosum* Beds (Figs. 2, 3) are distinguished in the basal part of the section (the Sinegorka Beds; Fig. 2, members 1–3). The assemblage consists of agglutinating species widespread in the lower Paleocene deposits of the North Pacific biogeographic province. *Haplophragmoides granulosum* Lipman is characteristic of the Danian Talitsa Horizon of the West Siberian Lowland (Podobina, 1975). *Silicosigmoilina longa* Serova was found in the Ivtygin Formation of the Koryak Upland and in the lower Paleocene Sinegorka Horizon of Sakhalin (Kalishevich *et al.*, 1981; Serova, 1987). *Haplophragmoides retrosepta* (Grzyb.) has a wide distribution in the Paleocene deposits of Western Europe (Hillebrandt, 1962), being also typical of the Sinegorka Horizon in Sakhalin and of the Ivtygin Formation of the Koryak Upland. *Ammobaculites expansus* Plummer is known from the Paleocene deposits of North America. *Haplophragmoides eggeri* Cushm. and *Asanospira glabra* (Cushm. et Waters) are widespread in the Maastrichtian and lower Paleogene deposits of North America, in the Sinegorka Horizon of the Naiba River basin (Sakhalin), and in the Ivtygin Formation of the Koryak Upland.

According to composition of foraminifers, the *Silicosigmoilina longa–Haplophragmoides granulosum* Beds are age analogues of the Sinegorka Beds of the upper Krasnoyarka Subformation in the Naiba River basin. As mentioned above, Vereshchagin *et al.* (1978) supposed that deposits analogous to the Sinegorka Beds occur in the Kril'on Peninsula. The foraminiferal analysis confirms this supposition. Characteristics of the foraminiferal assemblage from members 1–3 of the section studied indicates their early Paleocene age.

(II) The *Trochammina* Beds (Fig. 2, members 4, 5) correspond to the Snezhinka Formation. Abundant foraminifers infrequently constitute here the monotypic *Trochammina* assemblage. All four fractions macerated from sediments contain numerous deformed pyritized tests of single-type foraminifers unidentifiable at the species level. The tests may represent mineralized chitinous cover of the inner chamber surface of thin-walled species *Trochammina vitrea* Serova that inhabited tidal flats together with *Zostera*. Because of a poor preservation, these forms cannot be assigned to the mentioned species with confidence. Nevertheless, the similarity and probable identity with *Trochammina vitrea* may be used to outline stratigraphic distribution of this taxon in the Northwest Pacific region. It was found in the oyster beds of the Krasnopol'e Formation of the Avgustovka River section in Sakhalin and in the basal *Mytilus yokoyamai* Beds of the Snatol Formation of the Tochilin reference section in western Kamchatka (Serova, 1961). In the *Trochammina* Beds, the

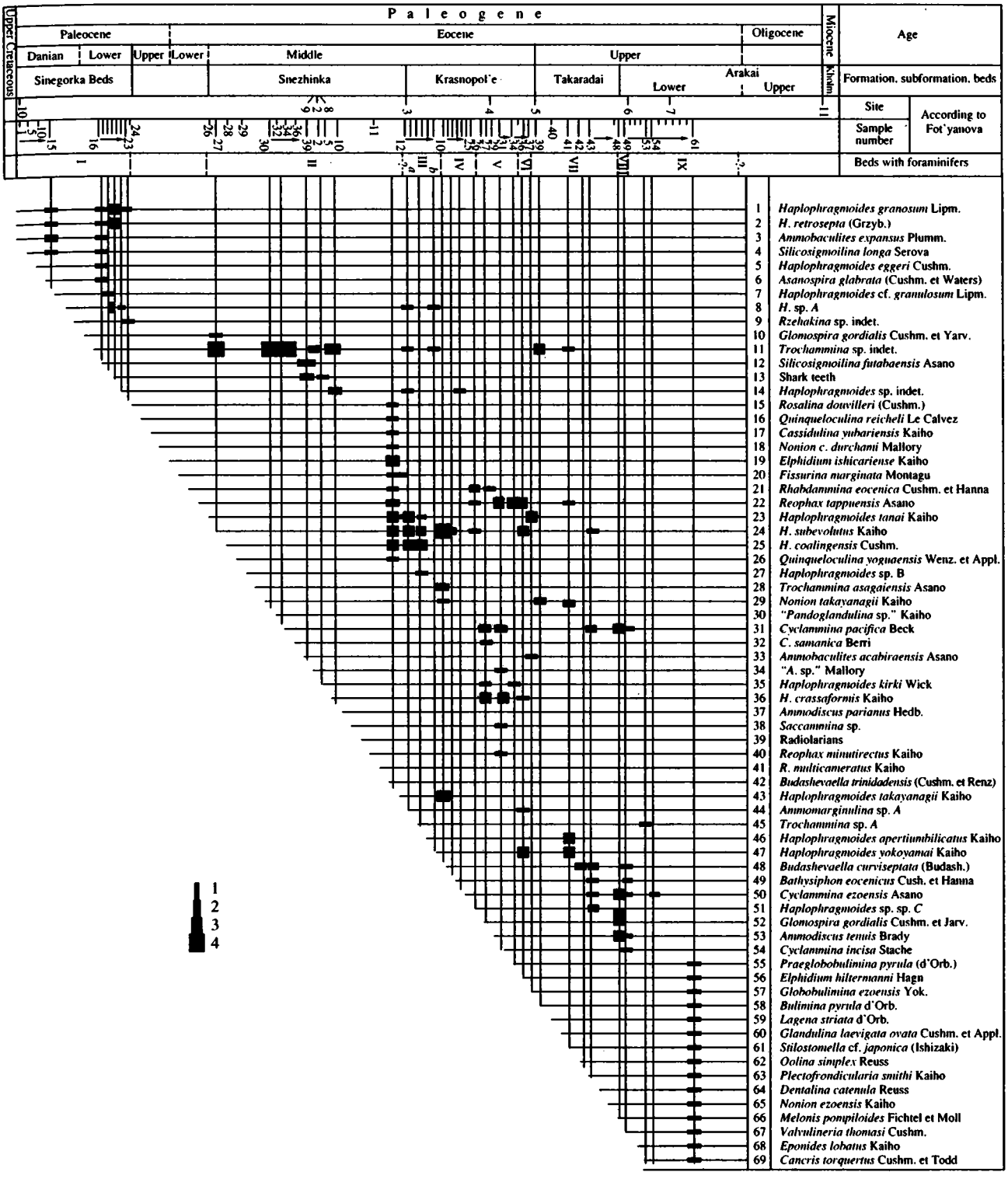


Fig. 3. Ranges of foraminifers in the Paleogene section at the Kitosiya River (according to Serova). Number of specimens in assemblages: (1) 1–5; (2) 5–10; (3) 10–25; (4) 25–100.

deformed tests of "*Trochammina* sp." are associated with single *Glomospira gordialis* Cushm. et Jarv., and *Silicosigmoilina futabaensis*.

When discussing age of the coal-bearing members 4 and 5 (*Trochammina* Beds), we should remember that the assemblage is impoverished and cannot serve as a

basis for the accurate dating of host deposits. An indirect evidence for the Eocene (most likely, middle Eocene) age of these members is the occurrence of a very diverse assemblage of benthic and planktonic foraminifers in the upper Paleocene–lower Eocene deposits of the Northwest Pacific (Serova, 1967, 1969; Krasheninnikov *et al.*, 1988). Assemblages of this kind have not been found in the carefully sampled Kitosiya River section. Accordingly we suggest the late Paleocene–early Eocene hiatus in sedimentation. The Snezhinka Formation of the Kitosiya River section may correspond to the marine lower part of the Snatol Formation of western Kamchatka and to the Krasnopol'e Formation bearing the *Trochammina* assemblage in the Avgustovka River section of Sakhalin. Therefore, the unit in question can correspond to the lower part of the middle Eocene. The floral assemblage of Member 5 (see the section "Characteristics of Macrofloras") is also of the middle Eocene age.

(IIIa) The *Elphidium ishikariense–E. asanoi* Beds (Figs. 2, 3) and their foraminiferal assemblage characterizes the basal part of the Krasnopol'e Formation (Member 7, samples 12 and 13[2]). The assemblage mainly consists of agglutinating foraminifers of the genus *Haplophragmoides* represented by *H. tanaii* Kaiho, *H. subevolutus* Kaiho, and *H. coalingensis* Cushm. Each of the species is represented by more than 100 specimens in every sample. The species *Reophax tappuensis* Asano is scarce. A characteristic feature of the assemblage is the occurrence of scarce but diverse secreting foraminifers *Quinqueloculina reicheli* Le Calvez, *Fissurina marginata* Montagu, *Cassidulina yubariensis* Kaiho, *Globocassidulina globosa* (Hantk.), *Nonion* aff. *durchami* Mall., *Elphidium asanoi* Kaiho, *E. ishikariense* Kaiho, and *Rosalina douvilleri* (Cushm.).

Species *Elphidium asanoi* Kaiho and *E. ishikariense* Kaiho present in the assemblage enable correlation between the lower part of the Krasnopol'e Formation and the Wakkanabe, Akabira and Hiragashi marine formations of the Ishikari Group in the Siomi area of Hokkaido (Kaiho, 1984). These are index-species of the of *Elphidium asanoi–Reophax tappuensis* (the Wakkanabe Formation) and *E. ishikariense–Bulimina yabei* (the Akabira and Hiragashi formations) zones, but they do not occur in the Poronai Formation. Being represented by single specimens in the lower part of the Krasnopol'e Formation, both species are abundant in Hokkaido, where their monotypic assemblages are characteristic of clayey facies (Kaiho, 1984).

Plentiful *Haplophragmoides tanaii* Kaiho and *H. subevolutus* Kaiho appear in the Paleogene section of the Ishikari area of Hokkaido beginning from lower beds of the Poronai Formation (Kaiho, 1984). Therefore, Beds IIIa could be correlated to Assemblage-Zone 5 of the Kaiho's scheme, as the indicated species are abundant in this zone. However, the characteristic secreting species, such as *Elphidium asanoi* Kaiho, *E. ishikariense* Kaiho, and others, do not pass into the Poronai Formation. Consequently, the correlation of the lower Krasnopol'e strata of the Kitosiya River section with the Wakkanabe, Akabira and Hiragashi formations of the Ishikari Group in the Ishikari area of Hokkaido seems to be quite correct.

Kaiho did not discuss age of the above formations. In earlier publications (Asano, 1958; Ujio and Wattanabe, 1960), the Ishikari Group, where marine deposits of the Wakkanabe, Akabira and Hiragashi formations are intercalated with continental ones, was correlated with the *Amphimorphina jenkinsi* Zone of the Kreyenhagen Formation of the Narizian Stage of California. On the basis of planktonic foraminifers, this stage is dated back to the latest middle Eocene (Mallory, 1959).

Other secreting foraminifers from Beds IIIa of the Krasnopol'e Formation are *Rosalina douvilleri* (Cushm.), *Quinqueloculina reicheli* Le Calvez, and *Nonion* aff. *durchami* Mallory, which are known from the middle Eocene deposits of western Europe (Murray and Wright, 1974). Thus, the lower Krasnopol'e strata bearing the *Elphidium asanoi–E. ishikariense* foraminiferal assemblage can be attributed to the upper part of the middle Eocene, if we take into consideration that, according to Kaiho, *Haplophragmoides tanaii* and *H. subevolutus* appear in abundance at the base of the Poronai Formation and are the index-species of the lower assemblage-zone of this formation. The stratigraphic inference based on the benthic foraminifer distribution is supported by nannoplankton found in lower beds of the Poronai Formation and characteristic of the upper part of CP14 Zone or to the lowermost part of CP15 Zone (Okada and Kaiho, 1992).

(IIIb) The *Haplophragmoides tanaii–H. subevolutus* Beds are distinguished in the upper part of Member 7 (samples 14–17[2]) of the Krasnopol'e Formation. The assemblage mostly consists of three plentiful species *Reophax tappuensis* Asano, *Haplophragmoides tanaii* Kaiho, and *H. subevolutus* Kaiho. Single specimens of agglutinating *Trochammina asagaiensis* Asano, *Trochammina* sp. and secreting *Quinqueloculina yeguaensis* Weinz. et Appl. occur sporadically. An analogous assemblage with more diverse secreting foraminifers characterizes the basal part of the Poronai Formation of the Mikasoporonai River section (Assemblage-Zone 5 according to Kaiho, 1984).

(IV) The *Haplophragmoides takayanagii–H. subevolutus* Kaiho Beds correspond to Member 8 (samples 19–25[2]) in the middle of the Krasnopol'e Formation. Two dominant species of the assemblage from the beds are the abundant *Haplophragmoides subevolutus* Kaiho passing from the underlying deposits and the newly appeared *H. takayanagii* Kaiho. They associate with single *Trochammina asagaiensis* Asano and *Nonion takayanagii* Kaiho.

(V) The *Haplophragmoides crassaformis–Cyclammina pacifica–Reophax multicameratus* Beds are

established in the upper part of the Krasnopol'e Formation (Member 9, samples 26–34)[2]. The assemblage of the beds comprises only agglutinating foraminifers. The dominant species are *Reophax tappuensis* Asano and the first appeared *Haplophragmoides crassaformis* Kaiho. Associated forms are single *Rhabdammina eocenica* (Cushm. et Hanna), first appeared *Reophax minutiformis* Kaiho, *Haplophragmoides subevolutus* Kaiho, *H. coalingensis* Cushm., *Budashevaella trinidadensis* (Cushm. et Jarv.), and *Ammobaculites akabiraensis* Asano. The characteristic feature is appearance of *Cyclammina* forms (*Cyclammina pacifica* Beck, *C. samanica* Berry) unknown in the underlying deposits of the section. In Hokkaido, the first occurrence *Cyclammina* species is recorded in basal beds of the Poronai Formation (Kaiho, 1984).

(VI) The *Haplophragmoides yokoyamai–Reophax tappuensis* Beds of the Krasnopol'e Formation (Member 9, samples 36 and 37)[2] yield only agglutinating foraminifers: dominant *Haplophragmoides yokoyamai* Kaiho (first occurrence), *Reophax tappuensis* Asano, *Haplophragmoides subevolutus* Kaiho, single *H. crassaformis* Kaiho, and *Ammomargulina* sp. The assemblage is much less diverse than the preceding one.

(VII) The *Haplophragmoides aperturiumbilicatus–H. yokoyamai* Beds are distinguished in Member 11 (Site 5, samples 41–43)[2] of the Takaradai Formation. As in the preceding assemblages, agglutinating foraminifers are most abundant and diverse here. *Reophax tappuense* Asano and *Cyclammina pacifica* Beck. are inherited from the underlying deposits. *Haplophragmoides aperturiumbilicatus* Kaiho, *H. yokoyamai* Kaiho, *Budashevaella curviseptata* (Budash.), and *Cyclammina incisa* appear at this level in abundance. The assemblage also contains single secreting species *Nonion takayanagii* Kaiho and *Fissurina* sp.

Foraminiferal beds VIII and IX characterize the lower Arakai Subformation.

(VIII) The *Haplophragmoides evolutus-Cyclammina ezoensis* Beds were established at the base of Member 12 (samples 48 and 49)[2], the lower Arakai Subformation. The 15-m-thick beds yielded a uniform assemblage of exceptionally agglutinating foraminifers. At this level, *Haplophragmoides evolutus* Kaiho is abundant. *Cyclammina ezoensis* Asano that is scarce below and *Cyclammina pacifica* Beck. are also frequent. This assemblage is the only one that is dominated by *Cyclammina* forms. Single *Budashevaella curviseptata* (Budash.), *Bathysiphon eocenicus* Cushm., *Glomospira gordialis* (Jones et Parker), *Ammodiscus tenuis* Brady, and *Cyclammina incisa* Stache are present. The assemblage of the beds is similar to that of the *Haplophragmoides subevolutus–Cyclammina pacifica* Assemblage-Zone of the Poronai Formation (Kaiho, 1984). Sal'nikov mapped the deposits bearing this assemblage as the Nikiforovka Beds.

(IX) The *Plectofrondiculria smithi–Globobulimina ezoensis* Beds characterize the upper part of the lower Arakai Subformation (Member 13, Site 6, Sample 61)[2]. Highly diverse secreting forms dominate in the assemblage consisting of *Haplophragmoides yokoyamai* Kaiho, "*Discammina* sp. A" Kaiho, *Lagena striata* d'Orb., *Fissurina marginata* Walker et Boys, *Oolina simplex* Reuss, *Dentalina catenula* Reuss, *Dentalina* cf. *jacksonensis* Cushm. et Appl. *D. lepidula* Schwager, *Stilostomella* cf. *japonica* (Ishizaki), *Glandulina laevigata ovata* Cushm. et Appl., *Plectofrondicularia smithi* Kaiho, *Cancris torquertus* Cushm. et Todd., *Valvulineria thomasi* Cushm. et Sim., *Eponides lobatus* Kaiho, *Nonion ezoensis* Kaiho, *Elphidium* aff. *sorahiense* Asano, *E. hiltermanni* Hagn, *Melonis pompiloides* (Fichtel et Moll), *Bulimina pyrula* d'Orb., and *Praeglobulimina pyrula* (d'Orb.).

In the northwestern Pacific biogeographic province, similar diverse assemblages of secreting foraminifers were found in the upper part of the Poronai Formation of Hokkaido (Assemblage-Zones 7–9 according to Kaiho, 1984), in the Peresheek Formation of Karaginskii Island (Serova *et al.*, 1975; Serova, 1986), and in the upper part of the Il'pinskii Group (the *Caucasina eocaenica kamchatica* Zone; Serova, 1976a, 1976b, 1983). In the Far East region, the Eocene–Oligocene boundary is placed between the Laparelam and Alugin formations in the Koryak Upland (the Il'pinskii Peninsula), between the Peresheek Formation (the upper part of the Mys Ton Formation, according to Gladenkov, 1972) and the Il'khatun Formation in the Karaginskii Island, and between the Poronai and Momijiama formations in Hokkaido.

In all these regional sections, this boundary is marked by sharp depletion of foraminiferal assemblages, which lost almost all secreting species plentiful in the upper Eocene deposits. Only few secreting species passed into the basal beds of the Momijiama, Il'khatun, and Alugin formations. Some agglutinating species also crossed the boundary to have a wide distribution in the Oligocene deposits.

Thus, the lower Arakai Subformation of the Kitosiya River section (Members 12 and 13) contains the diverse assemblage of foraminifers that can be dated back to the latest Eocene.

No foraminifers were found in the upper Arakai Subformation (Members 14 and 15), the Kholmsk and Nevel'sk formations. Only one sample from the Nevel'sk Formation (Site 12, Sample 81[2], Point *186*[4]) yielded *Cribroelphidium crassum* V. Kuzn. This species was described by V.N. Kuznetsova (see in Voloshinova *et al.*, 1970) from the Tumi Formation of the Shmidt Peninsula and also was found in Member 4 of the Arakai Formation of the Aral'skaya River section (Serova, 1983).

Paleogene foraminifers studied in the Kitosiya River section confirmed the previous conclusion based on investigations the Aral'skaya River section (Serova, 1983). Some Paleogene (Krasnopol'e, Takaradai, and Arakai) formations of southern Sakhalin have diachro-

nous boundaries and should be considered as lithofacies units differing in age range in different structural–facies zones rather than "formations" as they are defined in the Stratigraphic Code. So the Takaradai Formation of the Aral'slaya River section, which has been considered up to recently as the age analogue of the upper Eocene Poronai Formation of Hokkaido, has the early Oligocene age. The clayey Gennoishi Formation of the Aleksandrovsk area, which is correlative to the Takaradai Formation (*Geologiya...*, 1970), contains the late Oligocene assemblage of foraminifers (Serova, 1985).

The Arakai Formation of the Aral'skaya River section was accumulated in the late Oligocene time (Serova, 1983), whereas this formation of the Kitosiya River section, or at least its foraminifer-bearing lower part, is a subdivision of the latest Eocene age. The diachronism of boundaries of different lithofacies units, which are mapped as formations, seem to be natural if we take into consideration that the West Pacific sea transgression developed from the south to the north.

Mollusks. In the Kitosiya River section, mollusks are confined to the upper Paleogene deposits beginning from the Krasnopol'e Formation. Tuzov distinguished three molluscan assemblages (I, II, III in Figs. 2 and 4).

The first assemblage characterizes the lower part of the Krasnopol'e Formation (Member 7) and has almost the same stratigraphic range as the foraminiferal assemblage of Beds III. The molluscan assemblage consists of *Nuculana (N.)* sp., *Nuculana (B.) pulchrisinuosa* Clarc., *Yoldia takaradaensis* L. Krysht., *Yoldia* sp., and *Corbicula* sp. The only species passing into the overlying deposits is *Yoldia takaradaensis* L. Krysht.

The second assemblage characterizes the greater part of the Krasnopol'e Formation and the Takaradai Formation. It includes *Venericardia petvensis* Laut, *Mya* sp., *Nemocardium weavery* (And. et Mart.), *Macoma* sp., *Mya sachalinensis* L. Krysht., *Liocyma* sp., *Acila (Truncacila)* sp., *Megayoldia* cf. *nitida* Slod., *Periploma kariboensis* L. Krysht., *Ennucula* cf. *karatsuensis* (Khom.), *Acila* cf. *praesinnaensis* L. Krysht., and *Megayoldia cerussata* Slod.

Tuzov considers this assemblage as a single whole in contrast to opinion of other authors (Serova, Zharov, and Fot'yanova). Serova (1978) believes that evolution of marine faunas, particularly of foraminiferal assemblages, depended on evolutionary cycles of the basin. The most diverse foraminiferal assemblage characterizes the early transgressive part of the cycle (the Krasnopol'e Formation; Figs. 2–4; Beds IIIa and IIIb). Subsequently, in the middle part of the cycle (Beds IV and V of the Krasnopol'e Formation), the assemblage diversity decreases, finally to the complete disappearance of foraminifers in the late part of the cycle.

The analyzed distribution of mollusks in the section under consideration (Fig. 4) shows that they evolved in a similar manner: mollusks are most diverse in the lower part of the Krasnopol'e Formation and virtually absent in its upper part. Accordingly, Serova and Zharov divided the molluscan assemblage II into two subassemblages. The stratigraphic range of the lower subassemblage coincides with that of foraminiferal beds IV–VI, whereas the upper subassemblage characterizing the clayey Takaradai Formation corresponds in range to the foraminiferal beds VII. Tuzov and Gal'versen do not differentiate the subassemblages considering the molluscan assemblage II as indivisible. Their opinion is represented in Figs. 2 and 4.

The third assemblage is distinguished in the Arakai Formation. It consists of *Acila (Truncacila)* sp., *Megayoldia* cf. *nitida* Slod., *Periploma kariboensis* L. Krysht., *Nuculana (Borissia)* sp., *Malletia onorensis* Laut., *Laternula* cf. *besshoensis* (Yok.), *Acila (Truncacila)* cf. *osugii* (Nag. Et Huz.), *Venericardia* cf. *tokunagai* Yok., *Thracia condoni* Dall., and *Turcicula sachalinensis* L. Krysht. The first three of listed species passed into the Arakai Formation from the underlying Takaradai deposits. Other species have their first occurrence in the lower and upper Arakai subformations.

The fourth assemblage is characteristic of the Kholmsk Formation. It consists of two species unknown below: *Solemya tokunagai* Yok. and *Macoma* cf. *simizuensis* L. Krysht.

Characteristics of macrofloras. Plant remains were collected from two levels of the Kitosiya River section: from Member 5 of the Snezhinka Formation (Sites *8* and *9*)[2] and from basal beds of Member 8 of the Krasnopol'e Formation (Point *833* of Gal'versen, Fig. 2). Taxonomically similar floral assemblages from closely spaced sites *8* and *9* (approximately 20 m apart) can be considered as one assemblage characterizing the middle portion of the Snezhinka Formation.

The assemblage includes *Metasequoia occidentalis* (Newb.) Chaney, *Trochodendroides speciosa* (Ward (Berry), *Magnolia kryshtofovichi* Borsuk, *Platanus* sp. (two species, one of which has a peltate-based leaf), *Populus* cf. *celastrophylla* (Baik.) Sycheva, *Quercus* sp. (two species), *Fagus* sp., *Castanea longifolia* Borsuk, *Alnus* sp., *Alangium basitruncatum* (Oishi et Huzioka) Tanai, *Tilia* cf. *setadentata* Sycheva, and *"Cordia" japonica* Tanai.

Floral remains of the assemblage characterize the relatively large-leaved plants. *Trochodendroides speciosa* clearly dominates among them. Representatives of *Platanus* are of Paleocene–Eocene aspect. The most significant feature of the assemblage is presence of three Fagaceae genera. This feature is typical of the middle–late Eocene floras of Kamchatka, Sakhalin, and Japan (Budantsev, 1997; *Eotsen...*, 1991; Tanai, 1970, 1992; Fot'yanova, 1984a, 1984b).

As a possible equivalent of the assemblage, we can mention plant remains collected from the lower and middle Snezhinka subformations of the Snezhinka and Aral'skaya river sections. It is noteworthy that Grokho-

Eocene															Oligocene						Miocene	Series
Middle			Upper																			Subseries
Krasnopol'e									Takaradai					Arakai							Kholmsk	Formation
														Lower					Upper			Subformation
168	170		171				172		173					174		177			177a	178	182	183	Observation points
5, 6	1, 2	3	1	2	3	4	1	2	3	4	5	6	7	5	7	1	2	4		2	1		Samples
*	*						*	*				*											*Nuculana (N.)* sp.
*	*																						*Nuculana (B.) pulchrisinuosa* Clarc
*																							*Yoldia* sp.
*		*																					*Y. takaradaensis* L. Krisht.
*																							*Corbicula* sp.
		*		*																			*Venericardia petvensis* Laut.
		*	*			*																	*Mya* sp.
		*	*	*					*	*	*		*										*Nemocardium weavery* (And. et Mart.)
			*																				*Macoma* sp.
				*	*						*												*Mya sachalinensis* L. Krisht.
									*														*Liocyma* sp.
											*	*					*	*					*Acila (Truncacila)* sp.
											*				*								*Megayoldia* cf. *nitida* Slod.
											*	*							*		*		*Periploma kariboensis* L. Krisht.
													*										*Ennucula* cf. *karatsuensis* (Khom.)
													*										*Acila* cf. *praesinnaensis* L. Krisht.
													*										*Megayoldia cerussata* Slod.
														*									*Nuculana (Borissia)* sp.
														*	*	*							*Malletia onorensis* Laut.
														*									Laternula cf. *besshoensis* (Yok.)
															*								*Acila (Truncacila)* cf. *osugii* (Nag. et Huz.)
															*					*			*Venericardia cf. tokunagai* Yok.
																				*			*Thracia condoni* Dall.
																				*			*Turcicula sachalinensis* L. Krisht.
																						*	*Solemya tokunagai* Yok.
																						*	*Macoma* cf. *simizuensis* L. Krisht.

Fig. 4. Distribution of mollusks in the Paleogene deposits of the Kitosiya River section (according to Tusov's revision of materials by Savitskii and Margulis).

tova discovered a single palynological assemblage in these units (see in Sycheva, 1977).

The assemblage of the Kitosiya River section is most similar to that of the Tenneru and Harutori formations of the Urahoro Group in the Kushiro coal basin of Hokkaido (Fot'yanova and Serova, 1976). When fossil flora and biostratigraphy of the basin were studied better, the formations age changed from Oligocene (Tanai, 1970) to late Eocene (Kaiho, 1984). At present, we knew data on nannoplankton from the Sitakara Formation overlying the Yubetsu Formation in the Kushiro coal basin (Okada and Kaiho, 1992). The Sitakara Formation appears to be older than the nannoplankton zone CP14 (the upper part of the middle Eocene). Below the Sitakara Formation, there are downward-successive Yubetsu, Rushan (Tenneru) and Harutori continental formations, which are correctly supposed to be older than the upper half of the middle Eocene. Most probably, they are of the early–middle Eocene age.

On the contrary, Kodrul (1994) differentiated the lower and middle Snezhinka assemblages and correlated the lower one with that of the Noborikawa Formation of the Nemuro Group in eastern Hokkaido. Fot'yanova had the opportunity to examine the floral collection from the Noborikawa Formation in the National Science Museum of Tokyo. The examination revealed many plant remains of subtropical genesis. Most of them have small, narrow, and partly rigid leaves. This flora may be of the early Eocene age or may correspond to the transitional Paleocene–early Eocene climatic optimum. In addition, it shows a high endemism of genera and some families thus being distinct from coeval high-latitude floras of the Northeast Pacific. These features suggest that flora under consideration was connected with some other floral realm but not with the Boreal-Pacific one.

The second assemblage is typical of the middle part of the Krasnopol'e Formation. It is the monotypic, consisting of species *Macclintockia pugetensis* Wolfe. In the North Pacific, the acme-zone of this species spans the terminal middle–basal upper Eocene.

Palynological characteristics. As mentioned above, spore–pollen spectra were studied the same samples, which yielded foraminifers. Spectra from the lower, mainly continental deposits of the section are conventionally united into two palynological assemblages.

Assemblage I (Fig. 2) was distinguished in the Sinegorka Beds (Members 1–3, Site *10*, samples 8, 15–18, 21, 23)[2]. The assemblage includes pollen of coniferous taxa representing families Pinaceae and Taxodiaceae (33.6–48.4%), such as *Pinus* (*Pinus minutus* Zakl. included), less frequent *Picea, Cedrus, Tsuga,* Taxodiaceae, and *Scyadopitys.* Associated pollen is that of *Podocarpus* and *Ginkgo.* Grains of *Glassopollis,* widespread in the Jurassic deposits sporadically occur in the Eocene strata. The abundance rate of angiosperm pollen ranges from 15.6 to 33.1%, and 5.7–14.2% of taxa in this pollen group belong to Betulaceae. Broad-leaved plants are represented by *Corylus, Juglans, Pterocarya, Carya, Fraxinus, Acer, Tilia, Fagus, Quercites sparsus* Sam., *Quercus gracilis* Sam., *Castanopsis, Ulmoideipites planeraeformis* And., *U. tricostatus* And., *Hainnania, Liquidambar,* Hamamelidaceae and Ericaceae. The formal taxa include *Triatriopollenites rorubituites* Pfl., *T. confusus* Zakl., *Anacolosidites primidenius* Zakl., and *Orbiculapollis globosus* Chlon. (two last taxa are identified in the Sample 8 only). Spores correspond to frequent Polypodiaceae forms and *Leiotriletes,* and also to scarce *Sphagnum, Botrychium, Osmunda, Lygodium asper* Boldn., *L. japonicaeforme* E. Iv., and *Dicksonia arborescens.* The predominance of conifers and presence of relic form taxa (*Classopollis, Ulmoideipites planeraeformis* And., *Orbiculapollis globosus* Chlon., and others) do not contradict the opinion of Serova who attributed the Sinegorka Beds of the synonymous horizon to early Paleocene on the basis of foraminiferal stratigraphy.

Assemblage II characterizing the Snezhinka, Krasnopol'e and Takaradai formations is divided into two subassemblages (IIa and IIb). Subassemblage IIa characterizes Members 4 and 5 of the Snezhinka Formation (Site *10*, Samples 30, 32, 38; Site *2*, Sample 4)[2]. It is dominated by angiosperm taxa (36.4–42.0%) and demonstrates the increased abundance of Betulaceae (15.0–19.5%) and more diverse composition of broad-leaved plants (14.5–20.0%). Thermophilic forms among the latter are *Comptonia, Quercites sparsus, Castanea, Castanopsis, Fagus, Nyssa, Engelhardia, Juglans, Carya, Tilia, Hainnania, Ulmoideipites krempii, U. planeraeformis, Liliacidites, Hamamelidaceae, Liquidambar, Acer, Tilia. Tricolpites* and *Tricolporopollenites* of unknown systematic position are more frequent at this level. Spores and pollen of gymnosperms retain the same taxonomic composition, but their abundance decreases (20.8–32.7%). Subassemblage IIb characterizes Members 7 and 9 of the Krasnopol'e Formation (Site *3*, Samples 15, 15a, 19; Site *4*, Samples 29, 30, 33–35, 37)[2] and Member 11 of the Takaradai Formation (Site *5*, Samples 41, 47)[2]. Being similar in general to Subassemblage IIa, this one shows some distinctions. Abundance of spores varies throughout the section from 9.8 to 44.8%. The identified sporogenous taxa are Polypodiaceae, *Leiotriletes, Osmunda,* single *Sphagnum, Botrychium, Lycopodium, Selaginella, Lygodium japoniciforme* E. Iv., *L. asper* Boldn., and *Cyathea* associated with rare *Cingulatisporites eukirchensoides* Dele. et Sprum., *Pelletieria minutaestriata* Bolch., and *Aneimia, Gleichenia.*

Gymnosperm pollen increases in abundance from 13.5–30.4% at Sites *2* and *3* to 33.5–57.9% at Sites *4* and *5*. Spectrum of this group shows a higher proportion of large saccate pollen (*Picea* sec. *Eupicea*) and higher diversity of *Taxodium, Sequoia,* and *Cryptomeria* species. Abundance rate of *Pinus* and *Tsuga* pollen (*T. parva* Brutman and *T. torulosa* Zakl.) is low. Pollen grains of relic Cretaceous plants *Dacrydium, Podocar-*

pus, and *Ginkgo* are rare. Cretaceous forms *Classopollis* and *Gnetaceaepollenites* occur sporadically. Pollen of angiosperms (13.6–54.2%) includes that of Betulaceae (5.5–20.4%) and broad-leaved flowering plants (2.4–13.3%). Associated taxa of this group are scarce Late Cretaceous–Early Paleogene *Ulmoideipites planeraeformis* And., *U. tricostatus* And., *Corylopsis compacta* Lubom., *Quercites sparsus* Sam., *Anacolosiolites primigenius* Zakl., and *Elythranthe striatus* Couper. Form genera of the same age range are represented by rare *Aquilapollenites* and *Pistillipollenites mcgregori*. The latter is the indicative Eocene form in North America and Canada. In Sakhalin, this form is typical of the Paleocene–Eocene deposits.

Assemblage III is established in basal beds of the Lower Arakai Subformation (Member 12, Site *6*, Sample 52[2]; Point *174*[4]). It shows some difference as compared to assemblages from the underlying Snezhinka, Krasnopol'e and Takaradai formations. Spores, mainly those of Polypodiaceae and *Leiotriletes*, constitute 15.7% of the spectrum. Pollen of gymnosperms (56.5%) is that of *Taxodium, Sequoia, Glyptostrobus*, and *Sciadopitys* (family Taxodiaceae; 42.2%), and also of *Picea* and *Pinus* (11.2%). *Ginkgo* pollen is represented by single grains. Angiosperm pollen (20.4%) includes that of Betulaceae (11.8%) and of broad-leaved *Ulmus, Tilia*, and *Pterocarya* (3.9%). *Tricolpites striatellus* Chl., *Aquilapollenites quadrilobus* Rouse, and *Pistillipollenites mcgregori* have not been detected above the Eocene deposits. Scarce thermophilic miospores and Late Cretaceous–Early Paleogene *Aquilapollenites* and *Pistillipollenites* indicate the Eocene, most likely, the late Eocene age of the enclosing deposits.

SOME BIOSTRATIGRAPHIC REMARKS

As stated above, the lower Paleocene Sinegorka Beds of the synonymous horizon is the newly established stratigraphic unit of the Kril'on Peninsula. (The beds were recognized and dated on the basis of foraminifers and palynological spectra). It should be noted that in contrast to the stratotype of the Naiba River basin, these beds of the Kitosiya River section are mainly composed of terrigenous continental facies similar to those of the overlying Snezhinka Formation but intercalated with numerous marine interbeds.

Serova does not exclude that the succession of the Sinegorka Beds in the Kitosiya reference section is incomplete. The Naiba River section shows a gradual transition from Member 5 of the Krasnoyarka Formation to the Sinegorka Beds (Vereshchagin *et al.*, 1987; Kalishevich and Posyl'nyi, 1958; Kalishevich *et al.*, 1981). In opinion of Vereshchagin, Gal'versen, Voronin, and others who studied the Kitosiya River section these beds rest on evidently erosional surface of the fourth or even third member of the Upper Cretaceous Krasnoyarka Formation.

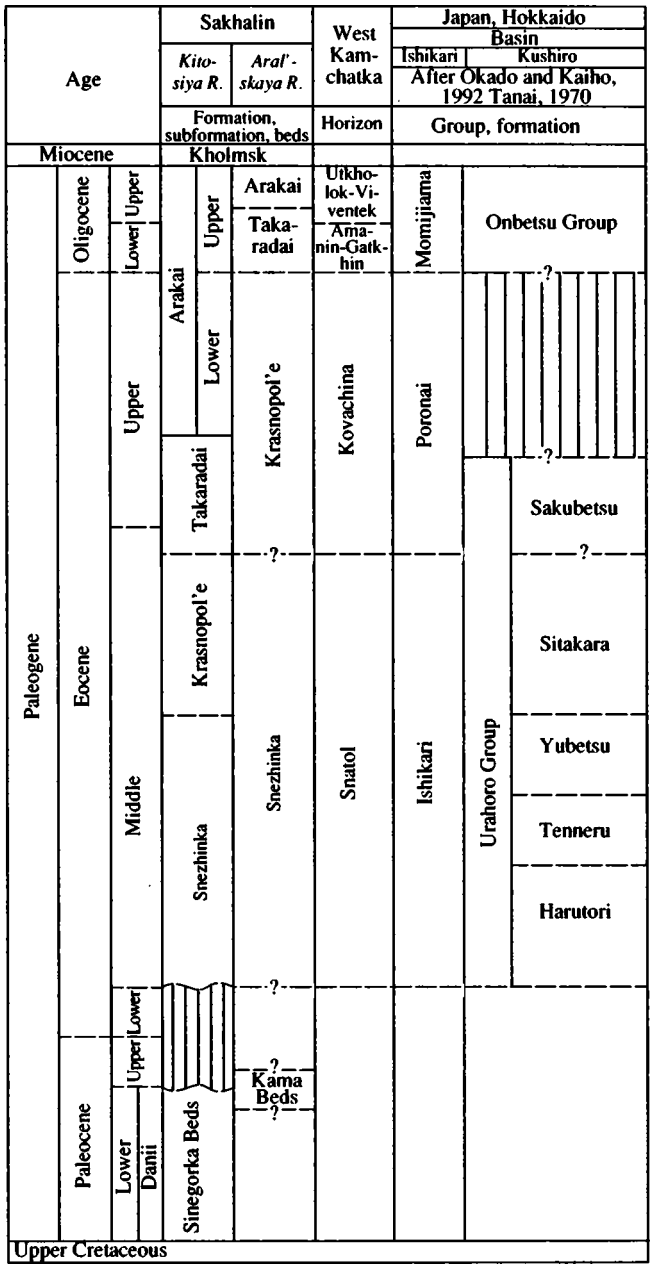


Fig. 5. Correlation between some Paleogene sections of Sakhalin, Kamchatka and Japan (compiled by Serova and Fot'yanova).

According to paleobotanical data, the overlying Snezhinka Formation is referred to the lower part of the middle Eocene. Thus, a significant (upper Paleocene–lower Eocene) gap can be suggested between the Sinegorka Beds and the Snezhinka Formation. The lower part of the Snezhinka Formation devoid of plant macrofossils (Member 4) may be however older than the middle Eocene. In addition, the Kitosiya River section is characterized by a sudden transition from the marine Sinegorka Beds, bearing foraminiferal assemblages of the relatively deep-water type, to the coastal-marine

and continental Snezhinka deposits with *Trochammina* assemblage of a very low diversity. This indicates an abrupt environmental change in the sedimentary basin. The problem concerning the scope of stratigraphic hiatus in the section under consideration needs to be studied further.

In contrast to other researchers who studied the section earlier (Zhidkova *et al.*, 1974; Il'ev *et al.*, 1977; Zakharova and Sheremet'eva, 1977), we accept a larger range of the Krasnopol'e Formation. First, all three members of this formation represent a single geological body mainly consisting of sandy and siltstone–sandy facies locally bearing coal seams. Second, the cyclic distribution of foraminifers (diverse assemblages at the beginning of the cycle and their absence at the end, at the top of the Krasnopol'e Formation), suggests, in opinion of Serova, that the formation as a whole corresponds to a single transgressive cycle in the basin development and should be dated back to the latest middle Eocene. Only two clayey members 10 and 11 constitute the thin Takaradai Formation, and Serova refers them to the upper Eocene on the basis of foraminiferal fauna. As distinct from other sections of Sakhalin, particularly from that of the Uglegorsk area (Serova, 1983), the lower subformation of the Arakai Formation is assigned in the Kitosiya River section to the upper Eocene as well with help of foraminiferal assemblages. In this section, the Eocene–Oligocene boundary is conventionally placed between the lower and upper subformations of the Arakai Formation (Fig. 5).

CONCLUSION

The comprehensive bed-by-bed study of Paleogene deposits and biota in the Kitosiya River basin suggest the following conclusions:

(1) The Paleogene section at the Kitosiya River is the only in the Kril'on Peninsula that has been studied in various biostratigraphic aspects. It fits all requirements for reference sections of structural–facies zones (*Zadachi...*, 1963).

(2) The succession of stratigraphic units in the section is used to establish their ages and relationships in the Kril'on subzone of the Southern structural–facies zone of Sakhalin.

(3) The foraminiferal *Silicosigmoilina longa–Haplophragmoides granulosum* assemblage characteristic of the lower Paleocene Sinegorka Beds of the stratotype section in the Naiba River basin (Kalishevich *et al.*, 1981; Serova, 1987) is discovered for the first time in basal Cenozoic deposits of the Kril'on Peninsula.

(4) The early–middle Eocene age of the Snezhinka Formation is established on the basis of paleobotanical evidence. The dating is supported by the monotypic *Trochammina* assemblage macerated from the formation.

(5) It is suggested that a stratigraphic hiatus spanning the upper Paleocene–lower Eocene interval separates the Sinegorka Beds and the Snezhinka Formation.

(6) The *Elphidium ishikariense–E. asanoi* Beds are recognized in the lower part of the Krasnopol'e Formation that corresponds to the beginning of the transgressive cycle. The beds are age analogues of the middle Eocene Wakkanabe, Akabira and Hiragashi formations of marine deposits in the Siomi area of Hokkaido. The thermophilic genus *Macclintockia (M. pugetensis)* Wolfe was found at the same level. Marine and continental biotas indicate that climate in the period under consideration was warmer than before and later.

(7) The foraminiferal analysis revealed diachronism of boundaries of the Krasnopol'e, Takaradai and Arakai formations. Their boundaries are older in the southern than in central Sakhalin.

(8) The comprehensive study of reference sections in different structural–facies zones, i.e., the detailed paleontological characterization of their stratigraphic units, is necessary to establish age ranges of "formations" and to elaborate the unified Paleogene stratigraphic scale for Sakhalin.

Reviewers M.A. Akhmet'ev and V.A. Krasheninnikov

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The Cenozoic of Sakhalin: Modern Stratigraphic Schemes and Correlation of Geological Events

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Abstract—Paleogene and Neogene stratigraphic schemes of Sakhalin are characterized. The schemes comprise from 11 to 12 regional stages and up to 50 formations. A calendar of various geological events (tectonic, volcanic, sedimentation, climatic, biotic, and others) is compiled on that basis. In particular, stages of different-scale changes in marine biotic assemblages are revealed and discriminated into major (15–17 Ma long) or subordinate and minor (2.0 and 0.3–1 Ma long). Several phases of the tectonic and volcanic activation, as well as periods of warming and cooling are distinguished. The analysis of geological events shows that many of them are synchronous to a certain extent, though it does not always happen that their causes and effects are understandable.

Key words: Paleogene, Neogene, Sakhalin, regional stages, zones, beds, correlation of events.

INTRODUCTION

Recently, a considerable attention has been focused on the study of transitional zones between oceans and continents. It is precisely in these zones where researchers managed to interpret many peculiar features of the Earth's history and understand, in particular, regularities in the formation of rock sequences, tectonic structures, and evolution of the past biotic communities.

The Sakhalin Island is one of significant geological objects that are now under study in the transitional zone between the Pacific Ocean and Asian continent. Widespread within the island are Cenozoic rocks, in which nearly 70 oil and gas fields, the largest in the Far East, are revealed that demonstrates the great practical value of the region.

As a geological structure, Sakhalin belongs to the geological block of the Sea of Okhotsk, the largest element of the West Pacific ocean–continent transitional zone, simultaneously representing a part of the Hokkaido–Sakhalin foldbelt. It is bounded by the East Sikhote-Alin volcanic belt on the west and by deep-sea basins on the east and south.

Three Cenozoic sedimentation basins: the North Sakhalin, West Sakhalin, and South Sakhalin one, are located in the Sakhalin and its shelf areas (Fig. 1). The basins emerged after folding and uplifting at the boundary between the Cretaceous and Paleogene under conditions of spreading and rifting. Coal-bearing formations were developed in Sakhalin at the beginning of the Paleogene. Marine conditions set up there in the Middle–Late Eocene when the sedimentary cover of the Sea of Okhotsk began to form. The sea transgression of that time left its records on eastern and northern margins of the Sea of Okhotsk (on the Kamchatka Peninsula and the Koryak Upland). The Oligocene subsidence involved the maximum area of the island, and Oligocene marine clayey–siliceous deposits are present even in the East Sakhalin and Susunai uplands. The bulk of sediments accumulated in the above-mentioned basins during the Miocene. The coal-bearing formation signifying a relative uplifting of the island accumulated during the terminal early Miocene–initial middle Miocene. The major transgression of the middle Miocene time expanded over many Sakhalin areas. A compression regime replaced that of crustal extension in the latest Miocene–earliest Pliocene, when a system of regional slip and reversed strike-slip faults was formed together with associated folds. The principal phase of folding and uplifts that created the modern structure of the region occurred in the latest Pliocene–earliest Quaternary.

Data on the Cenozoic geology and, particularly, on stratigraphy of the Sakhalin Island were excellently generalized at the beginning of the 1970s (*Geologiya...*, 1970). Thirty years have passed since that time. The study of Cenozoic rocks in Sakhalin significantly advanced in recent years: new geological maps have been compiled and reviews on two key sections of the Makarov and Shmidt regions have been issued (Arkhipova *et al.*, 1992; Gladenkov *et al.*, 1999). The regional stratigraphic scheme for the Cenozoic was published in 1998 (*Resheniya...*, 1998). Nevertheless, it is important now to reconsider and generalize data on stratigraphy, lithology, seismostratigraphy, oil and gas prospects, and paleogeography on the basis of new materials obtained after the late 1960s–early 1970s.

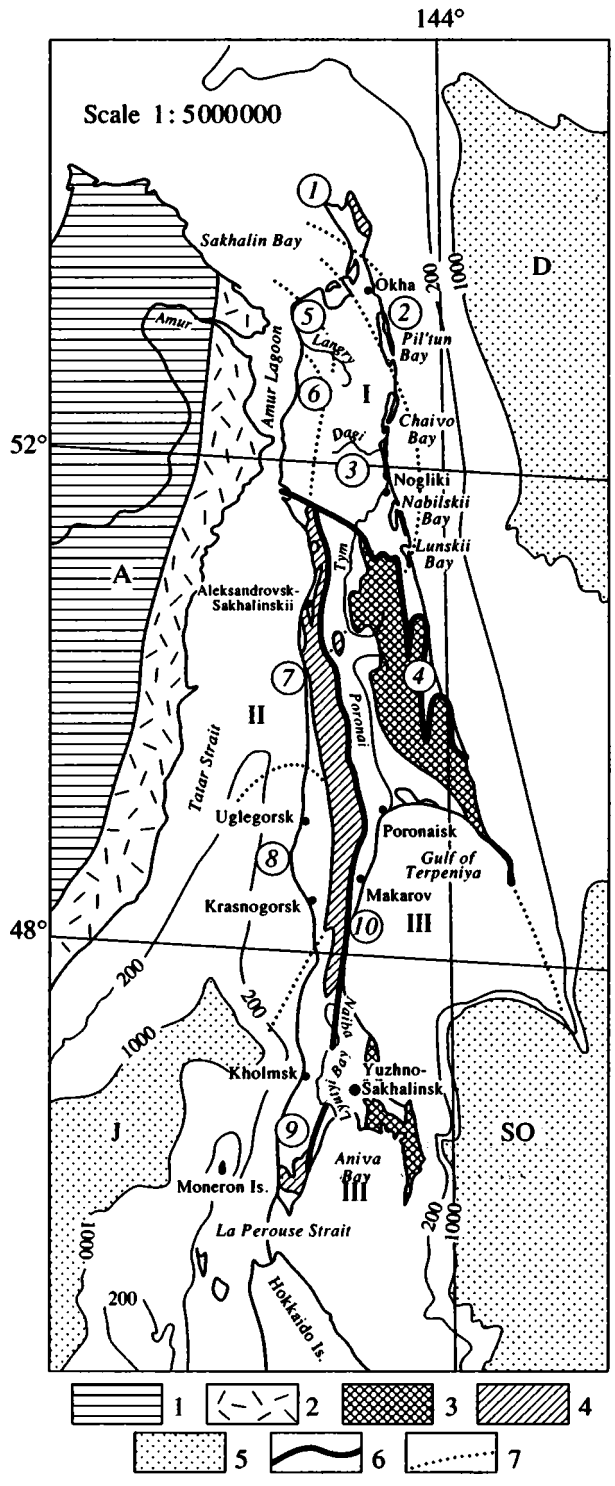

Fig. 1. Major Cenozoic structures and sedimentation basins of Sakhalin: (1) Sikhote-Alin folded zone (A); (2) East Sikhote-Alin volcanic belt; (3–4) the Hokkaido–Sakhalin folded zone with exposures of the Paleozoic–Mesozoic (3) and Upper Cretaceous (4) basement rocks; (5) Deryugin (D), South Okhotsk (SO), and Japan (J) deep-sea basins; (6) boundaries of Cenozoic sedimentation basins (structural-lithologic zones) of Sakhalin; (7) boundaries of structural–facies zones (regions). Sedimentary basins: (I) North Sakhalin, (II) West Sakhalin, (III) South Sakhalin. Reference sections (encircled numbers): *1*, Shmidt Peninsula; *2*, Okha; *3*, Dagi; *4*, Pogranichnyi; *5*, Langry; *6*, Engizpal; *7*, Aleksandrovsk; *8*, Uglegorsk–Chekhov; *9*, Kholmsk–Nevel'sk; *10*, Makarov.

Such an attempt was made by Gladenkov, Bazhenov, Grechin, Margulis, and Salnikov whose work is now under preparation for publication. The work is aimed to characterize ten key sections of the Cenozoic, their lithology and paleontology, seismostratigraphic data on the shelf zone, oil and gas showings, and also to give an analysis of paleogeographic environments and geological events. In this article, I would like to touch only upon three aspects among all mentioned above: first, to consider general problems that emerge when distinguishing stratigraphic units of different rank (horizons, zones, etc.) and compiling the Cenozoic correlation scheme; second, to analyze changes in the Cenozoic biotic communities; and third, to correlate different geological events on a unified stratigraphic basis and to compile the general calendar of those events.

HORIZONS, ZONES, AND BEDS: THEIR PALEONTOLOGICAL SUBSTANTIATION AND AGE

A long practice of geological exploration in Sakhalin showed applicability of local and regional stratigraphic units for subdivision of Cenozoic sedimentary sequences. The commonly used local unit is a formation (divisible into subformations and members) often similar in its essence to the same stratigraphic unit in understanding of American geologists. A horizon, or a regional stage, is the basic unit of the regional status. Reflecting particular stages of the geological history of the region, horizons represent the units, to which numerous formations (about 50) distinguishable in different structural–facies zones of the island and differing from each other in the thickness and lithological aspects (terrigenous and volcanogenic sequences) are referred to. I should mention here that most sections of the Cenozoic rocks in Sakhalin are up to 5–6 and even 11–12 km thick.

In total, 11 to 12 horizons are distinguishable within the Cenozoic, more exactly, Paleogene–Neogene sequence of Sakhalin. Sometimes subregional units (horizons, subhorizons), which with their similar ages may have different names or different ranges, are recognized in northern Sakhalin, on the one hand, and the southern and western areas of the island, on the other. However, such units are always correlative with general regional stages of Sakhalin. For instance, the Okobykai, Lower Nutovo, and Upper Nutovo horizons distinguished in the middle Miocene–Pliocene in northern Sakhalin correspond in fact to regional stages of the same name characterizing the whole Sakhalin area.

Insouthern Sakhalin, they correlate with the Kurashiki (with two subhorizons) and Maruyama horizons.

It does not always happen that the Sakhalin horizons are distinguished properly in terms of methodology. Unlike Paleogene horizons, which are discriminated within a single stratotype locality (western Sakhalin), where one can observe a clear succession of all units and their boundaries, the Neogene units overlying the Kholmsk Horizon are often distinguished inadequately, because their type sections are hardly accessible and usually discontinuous, and the horizons include formations and sequences studied in different places, and so on. In order to distinguish Neogene horizons more correctly, it would be appropriate to use the accessible and exposed section of the Makarov region (southeastern Sakhalin), where the succession of beds is well seen and all the strata have perfect paleontological characteristics.

Currently, the stratigraphic scheme of Cenozoic rocks of Sakhalin consists of the following horizons (from the base upward):

(1) Snezhinka, Paleocene–middle (?) Eocene; (2) Krasnopol'e, middle Eocene; (3) Shebunino, upper Eocene; (4) Arakai and Kholmsk, Oligocene; (5) Uina, the basal lower Miocene; (6) Dagi, lower Miocene–lowest middle Miocene; (7) Okobykai, middle Miocene–basal upper Miocene; (8) Lower Nutovo, upper Miocene; (9) Upper Nutovo, terminal upper Miocene–basal Pliocene; (10) Pomyr, middle–upper Pliocene. The new Deryugin horizon of the uppermost Pliocene–Quaternary usable sometimes in recent years has not been officially authorized; sediments of the outer shelf of northern Sakhalin are assigned to this unit.

The analysis reveals irregular time ranges of the horizons. If the Snezhinka Horizon spans c.a. 20 (?) m.y., ranges of the Dagi, Arakai, and Okobykai horizons are 7.0, 5.0, and 5.0 m.y. respectively, the Upper Nutovo Horizon is 3.0 m.y. long, and the Pomyr Horizon lasted about 1.0 m.y. This may indicate that the Snezhinka interval is subdivided yet insufficiently.

Paleontological characteristics of the horizons are different. The lower, Snezhinka Horizon is characterized by flora, whereas others mainly by mollusks and benthic foraminifers, partly by leaf flora, spores, and pollen. Diatoms play a highly important role in characterization of Oligocene-Pliocene deposits. Nanoplankton and dinocysts have been revealed at some Eocene levels in recent years. Radiolarians are found in places. The correlation of the fossil assemblages with those from Paleogene and Neogene strata of Kamchatka, Japan, and North America, as well as assemblages found in bottom sediments of adjacent areas of the Pacific Ocean, allow us to compare the Sakhalin horizons with units of the general scale and to establish their age. Age levels are reliably defined on the basis of zonal planktonic assemblages (primarily diatoms). Many correlations can be substantiated using benthic fossil groups as well.

A number of specialists, among which there are great authorities on the subject, took part in studying paleontological materials from Cenozoic strata of Sakhalin. Mollusks were investigated by L.V. Krishtofovich, F.G. Lautenshleger, I.P. Khomenko, L.S. Zhidkova, V.O. Savitskii, Yu.B. Gladenkov, T.G. Kalishevich, I.N. Kuzina, and others, foraminifers were studied by N.A. Voloshinova, V.N. Kuznetsova, L.S. Leonenko, L.I. Mitrofanova, M.Ya. Serova, and others, diatoms by A.I. Moiseeva, V.P. Boldyreva, T.V. Oreshkina, A.Yu. Gladenkov, and others, spore-and-pollen spectra by L.A. Toboyakova, N.Ya. Brutman, A.D. Arkhipova, N.M. Grokhotova, and others, macroflora by A.N. Krishtofovich, V.A. Krasilin, L.I. Fot'yanova, O.A. Sycheva, T.V. Kodrul, and others. The list is incomplete of course. It is significant that all these researchers contributed much to the comprehensive paleontological substantiation of the Sakhalin horizons, characterizing them on the basis of different groups of the fauna and flora.

Several monographs devoted to the Cenozoic biota of Sakhalin have been published. Two of them deal with mollusks of Sakhalin (Krishtofovich and Il'ina, 1954; Krishtofovich, 1964), one book with foraminifers (Voloshinova, *et al.*, 1971), and another one with the Oligocene–Pliocene stratigraphy and paleontology of the Makarov region (Arkhipova *et al.*, 1992) and the Shmidt Peninsula (Gladenkov *et al.*, 1999). Some monographs are dedicated to the flora of the region (Fot'yanova, 1988; Kodrul, 1994). On the one hand, the investigation level cannot be taken for excellent, as many sections and assemblages remained insufficiently characterized (for instance, the comprehensive characteristic of fauna from sections of southwestern Sakhalin is still unknown). On the other hand, a valuable data base on biostratigraphy is already created, and it can be used for the general correlation and paleogeographic reconstructions.

The books of the early 1970s dedicated to biofacies peculiarities of paleobasins of Sakhalin (e.g., Zhidkova *et al.*, 1974) exemplify fruitful attempts of paleontological data generalization (first of all, those on mollusks). Important paleontological characteristics of some sections from different regions of Sakhalin are also presented in numerous articles. I think that readers deeply interested in these problems know all these publications, and here I would like to focus attention on some peculiarities of paleontological characteristics of the horizons and on some debatable points concerning their age substantiation.

I should point out first that it became a common practice for stratigraphers to operate with smaller stratigraphic units, e.g., with zones and beds with fauna. For instance, more than 15 zones are distinguished on the basis of diatoms within the Oligocene-Pliocene sequence. Precisely these zones enabled the highly

accurate correlation of Sakhalin strata with the oceanic scale. Unfortunately, the zones are distinguished not in all sections, and their succession is not always continuous. Nevertheless, in regions where sections are studied in detail, the detailed subdivisions, i.e., local zones or beds with fauna, can be distinguished, in particular, on the basis of mollusks and foraminifers. Beds with mollusks, which are recognized in the section of the Shmidt Peninsula, are good examples. Nine beds with indicative molluscan assemblages are distinguished within the Machigar and Tumi formations corresponding mainly to the Oligocene–early Miocene time range. It is remarkable that some of the beds are recognized in other regions as well, but the comprehensive bed-by-bed study of fauna is unfinished, and accordingly, it is difficult to trace all the beds (or their age analogs) throughout the whole territory.

Presented below is brief characteristic of each horizon (from the base upward).

The Snezhinka Horizon (the Paleocene–lower half of the Eocene) is composed of continental, often carbonaceous deposits up to 1500 m thick. It is characterized only by plant fossils known under the name the "early Due flora." Common species encountered in the horizon are *Fagus paucinervis, Trochodendroides arctica, Credneria grewiopsoides*, and others. Two palynozones of the Muz'ma spore-and-pollen assemblage (SPA), namely the lower *Alnipollensis anacolosidites* and upper *Lygodium japoniciforme–Ulmoideipites* zones, are distinguished as well. The recent tendency is to subdivide the Snezhinka flora into two parts separated by a hiatus (Kodrul, 1994): the lower (the Kamchik beds) coniferous–broad-leaved assemblage corresponding to the early stage of the Paleocene–Eocene warming, and the upper one (the middle (?) Eocene) of coniferous–small-leaved plant remains. The suggested hiatus should be verified by special works in the future. Rare foraminifers (*Ammodiscus macitentus* and others) are found in the upper part of the horizon.

The Krasnopol'e Horizon (the middle Eocene) is mainly composed of marine sandstones up to 1500 m thick. Characteristic, relatively thermophilic mollusks appear here: *Nuculana snatolensis, Ostrea agnevoensis, Pitar californiana*, and others. Reported foraminifers are *Haplophragmoides asagaensis, Pseudoelphidiella californica*, and others.

The Shebunino Horizon (the upper Eocene) comprises siltstone–mudstone and sandstone strata up to 1000 m thick. Mollusks are represented by *Yoldia takaradaiensis, Malletia poronaica, Nemocardium ezoense*, and others, many of which are typical of adjacent areas as well. Also characteristic is foraminiferal assemblage with *Plectofrondicularia packardi* and others known from many areas from Japan and Kamchatka to California.

The Arakai Horizon (the Oligocene) consists of tuffaceous sandstone, psephitic tuff and siltstone beds up to 1000 m thick. The molluscan assemblage very characteristic of it comprises *Papyridea matschigarica, Yoldia caudata, Portlandella watasei, Modiolus matchgarensis*, and others. Seven units, namely the *Papyridea harrimani, Nemocardium iwakiense, Trachycardium kinsimarae, Acilia oyamadensis, Glycymeris nakosoensis, Delectopecten watanabei*, and *Cardiomya majanatschensis* beds (are distinguished from the base upward within the horizon in the Shmidt Peninsula. These beds reflect changes of paleobiota in response to evolution of mollusks and to variations in paleogeographic and paleoecological conditions as well. Provided that changes are similar in different regions of Sakhalin, the beds can be recognized in many localities, but in case of distinction, they should be interpreted in a more generalized mode.

In recent years, a composite Arakai–Kholmsk Horizon is usually referred to the Oligocene. I reckon the horizon should be subdivided into the Arakai and Kholmsk parts, because the Kholmsk part (Horizon) comprising sandy–clayey to siliceous rocks up to 1200 m thick has its own paleontological characteristics. Typical among its mollusks are *Acila kholmskiensis, A. gettysburgensis, Nuculana crassatelloides, N. pennula, Cyclocardia tokunagai*, and others. Foraminifers of the Arakai level are represented by the assemblage with *Haplophragmoides laminatus, Melonis shimokinensis*, and others. The Kholmsk Horizon yields *Pseudoelphidiella subcarinata, Cribroelphidium crassum*, and other forms. Comparing the fauna lists, we can see that both horizons have similar taxa as well.

Data on diatoms appear to be very interesting among new biostratigraphic materials from Oligocene deposits. First, these data enabled correlation of separate sections with the general scale. Second, their fossil assemblages are better dated now that increases their correlation potential. The Oligocene diatom assemblages were recognized in the Pilenga Formation of the Pogranichnyi region (data of V.P. Boldyreva, T.V. Oreshkina, and A.Yu. Gladenkov). The *Rocella gelida* Zone was distinguished there in the upper part of the formation. The new data on the zonal stratigraphy of the Oligocene and lower Miocene is of a great importance for Sakhalin, since only middle Miocene and younger diatom zones were used before in stratigraphic practice here.

It should be noted that the Paleogene nannoplankton was recently revealed for the first time in southwestern Sakhalin (Gladenkov and Shcherbinina, 2001). For instance, *Coccolithus formosus* (Kamptner) (the lowermost Eocene) was found in the Krasnopol'e Horizon exposed along the Aralskaya River, whereas *Reticulofenestra dictyoda* (Deflanre) and *Dictyococcites bisectus* (Hay, Mohler, Wide) Bukry and Percival of the middle Eocene–basal upper Eocene were recognized in the Takaradai Formation exposures along the Tomarinka River. Species *Dictyococcites bisectus* and *Cyclicargolithus floridanus* (Roth and Hay) Bukry (the upper part of the Eocene–Upper Oligocene) were iden-

tified in the Arakai Formation. These fossils are important for the age substantiation of Paleogene strata in Sakhalin.

The Uina Horizon (the basal lower Miocene) represented by sandstone, siltstone, and siliceous rocks (up to 1500 m thick) is characterized by molluscan species *Nuculana alferovi, Periploma altarata, Mytilus ocholensis*, and others. Foraminifers recognized here are *Cyclammina pilvoensis, Pseudoelphidiella subcarinata*, and others.

The Dagi Horizon (its uppermost beds excluded) belongs to the early Miocene as well. It comprises in different regions coal measures (first of all, those of the Verkhnii Due Formation), shallow-water and relatively deep-water deposits, as well as volcano-sedimentary and volcanogenic rocks (the Nevel'sk and Chekhov formations). The horizon is up 1500–2000 m thick in total.

Typical mollusks of the unit are *Nuculana nabiliana, Acilana tokunagai, Mizuhopecten subyessoensis, Delectopecten pedroanus, Solemya dalli*, and others; characteristic foraminifers are represented by *Elphidiella simpex. Milliamina complanata, Pseudoelphidiella problematica*, and others. The Dagi Horizon of northern Sakhalin (the Kolendo beds) also yields *Ammonia tochigiensis, Siphogenerina*, and others. Its flora is represented by *Ginkgo adiantoides, Metasequoia occidentalis, Ulmus longifolia, Hemitrapa borealis*, and others (this is so-called Verkhnii Due flora). Diatoms are recognized in individual formations. For instance, *Thalassiosira praefraga, Thalassiosira fraga* and some other zones now attributed to the lower Miocene are revealed in the Borsk Formation of the Pogranichnyi region.

The upper part of the Dagi Horizon yields a characteristic, relatively thermophilic assemblage of paleobiota characterizing the initial middle Miocene. In my opinion, this characteristic level should be recognized as a separate horizon. Representative mollusks *Anadara watanabei, Glycymeris cishuensis, Dosinia tugaruana, Spisula haboroensis*, and others are identified at this level in the Ausi Formation and its analogs. Foraminifers *Haplophragmoides impressus, Glandulina nipponica, Asteroammonium takanabensis*, and others are encountered here as well. The level coincides with the base of the diatom *Denticulopsis lauta* Zone.

The Okobykai Horizon (the middle–basal upper Miocene) is composed of siliceous mudstone, diatomite, and sandstone beds up to 2000 m thick. Among representative mollusks, there are *Nucula psjakauphensis, Nuculana schmidti, malletia kurasiensis, Delectopecten pedroanus*, and others. Two foraminiferal zones are distinguished here: those of *Budashevella laevigata* and *Haplophragmoides renzi*. The horizon corresponds to the *D. lauta* and *D. lauta–D. hustedtii* diatom zones.

The Nutovo Horizon (the upper Miocene–lowermost Pliocene) is divided into lower (the upper Miocene) and upper (mainly the lowermost Pliocene) subhorizons (sometimes distinguished as Nizhnii and Verkhnii Nutovo horizons). The horizon comprises different lithological facies: shallow-water sandy–clayey deposits intercalated with relatively deep-water siltstone and diatomite beds, and also with subcontinental sandy–carbonaceous strata. Its total thickness is 3000–3500 m.

The lower Nutovo Subhorizon spans the *Denticulopsis hustedtii* and partially *Neodenticula kamchatica* diatom zones. The zones correlate with the *Haplophragmoides impressus, Uvigerina subperegina, Islandiella laticamerata* and other foraminiferal zones. Mollusks *Nuculana praerobai, N. vengerensis, Neptunea magna stantoni*, and others are found here as well.

The upper Nutovo Subhorizon corresponds to the greater part of the diatom *Neodenticula kamtschatica* Zone. Typical foraminiferal assemblages are those of the *Islandiella laticamerata* and *Retroelphidium clavatum* zones. The latter correspond to beds with mollusks *Fortipecten takahashii, Nuculana majamphensis, Spisula voyi*, and others.

The Pomyr Horizon (the terminal Pliocene) is usually represented by shallow-water and continental strata up to 1500 m thick. Two diatom zones of *N. kamtschatica–N. koizumii* and *N. koizumii* correspond to it. Guide forms of foraminifers are *Retroelphidium clavatum, Cribroelphidium subacticum*, and others. Typical mollusks of the unit are *Astarte alaskensis, Chlamys tanassevitschi*, and others.

To extend this brief description, I should mention first that most horizons include different lithofacies (different but coeval formations). This is quite understandable, because shallow- and relatively deep-water sedimentation settings coexisted at each evolutionary stage of different sea basins that is evident from lithology of sediments and from composition of biotic communities. Continental sedimentation conditions could exist simultaneously. Clearly it would be well to reveal successive and widespread benthic and planktonic assemblages in each horizon, which corresponded to each facies—tidal, upper and lower subtidal, and bathyal. Unfortunately, such "sets" of species and genera of mollusks and foraminifers changing with time are far from being completely known in detail. Although, some of their general features are already established in terms of zoogeographic, vertical, and trophic zonation of biocoenoses (Zhidkova *et al.*, 1974). The foregoing lists of flora and fauna from separate horizons characterizes first of all the representative and dominant taxa.

Another particular emphasis should be placed upon different ranges of established diatom zones. Fossils collected in Sakhalin earlier are used to recognize zones in the scale that was elaborated in the 1970s on the basis of work by Koizumi (1975) with later contributions by other researchers. These were mainly zones of the middle–upper Miocene and Pliocene. In the 1990s, the diatom zonations elaborated in detail

extended down to the early Miocene and Oligocene (Gladenkov and Barron, 1995; Barron and Gladenkov, 1995). However, they are useful only for interpretation of new materials of the 1980s (e.g., of new data on Cenozoic deposits of the Shmidt Peninsula and Pogranichnyi region). Nevertheless, this should not be a serious handicap to subdivision of sequences studied earlier, as the previous and new zonations are reliably correlated.

My third remark concerns such units as "beds with fauna." Their distinguishing is of great importance in the methodological and practical respect, and pioneering attempts to recognize such beds in Sakhalin deserve a special attention. Actually, I mean a further detailing of stratigraphic scales at the level of "infrazones" necessary for the large-scale mapping and a high-precision interpretation of drilling results (Gladenkov, 1995).

The distinguishing of beds leans upon a comprehensive study of representative fossil assemblages collected layer-by-layer. For instance, the collection of mollusks used for subdividing the Oligocene–Pliocene of the Shmidt Peninsula includes more than 2000 specimens of more than 170 species. By the comparative analysis of molluscan assemblages from different beds, the following parameters have been considered: (1) the total number of species, (2) the number of species appearing and disappearing in each bed, (3) the number of extinct and recent forms, (4) the number of subtropical and boreal species, (5) the number of endemic species, (6) the number of dominant forms, etc. The comparison of these parameters enabled discrimination of beds with fauna. At present, when beds of that sort are distinguished in many regions of Kamchatka and, after recent "experimental" works in the Shmidt Peninsula as well, the beds may find the proper place in the Cenozoic stratigraphic schemes of other Sakhalin regions after the appropriate studies. A layer-by-layer study of mollusk assemblages from sections of northern Sakhalin (the Shmidt Peninsula) revealed 17 successive beds having definite paleontological characteristics (see Gladenkov *et al.*, 1999 for details). The stratigraphic succession of distinguished units from the base upward is as follows: the Machigar Formation comprising the *Mytilus littoralis–Corbicula sitakaraensis, Papyridea harrimani–Macoma sejugata, Nemocardium iwakiense–Thracia schmidti, Trachycardium kinsimarae–Yoldia matschigarica, Acila oyamadensis*, and *Glycymeris nakosoensis* beds; the Tumi Formation with the *Delectopecten watanabei–Nucuclana tuniensis, Cardiomya majanatschensis*, and *Chlamys donmilleri–Mytilus ochotensis* beds; the Pil'vo, Kaskad, and Vengeri formations with the *Delectopecten pedroanus–Nuculana schmidti, Acilana tokunagai–"Yoldia" multidentata, Crenomytilus expansus–Panomya informedia*, and *Conchocele disjuncta ochotica–C. alta* beds; the Mayamraf and Matituki formations with the *Nucuclana majamraphensis–Nucula psjakauphensis*, and *Potamocorbula amorensis* beds; and the Pomyr Formation incorporating the *Chlamys tanassevitschi–Astarte alaskensis* and *Macoma calcarea* beds. Some groups of these beds are traceable in other regions of Sakhalin, although the inadequate data on some sections often inhibits their correlation.

Unfortunately, such a detailed recognition of beds with foraminifers is rare nowadays practice, although their generalization on the basis of particular sections may give good results. Three persistent foraminiferal assemblages are known for the Paleogene (to be more exact, for the Eocene–Oligocene) of Sakhalin, and they help to distinguish corresponding beds in a series of sections. The fairy detailed succession of beds (8 to 10) is distinguished within the Miocene interval. I believe that in future, discriminating beds with foraminifers and using the unified technique and comprehensive analysis of fossil assemblages, we will be able to make the correlation schemes more reliable.

I do not touch here the issue of spore–pollen assemblages (SPA) and palynozones, which, to be sure, are of great importance in stratigraphy but less detailed as compared to beds and zones distinguished on the basis of sea organisms. Nevertheless, one should remember about their clear stratigraphic position and their potential for stratigraphic subdivisions. For instance, the Muz'ma SPA characterizes the Eocene, the Berezovyi SPA the Oligocene–lower Miocene, the Langry, Okha, and Ossoi I assemblages the Miocene, and the Ossoi II SPA the upper Miocene–Pliocene. It might be well to point out that palynozones very often help to distinguish those levels in sections which mark paleoclimatic fluctuations (warming and cooling episodes) thus representing additional correlation markers (Brutman, 1986).

Lastly, the fourth remark concerns the correlation of biostratigraphic units distinguished on the basis of different paleontological groups. Such a correlation controls positions of boundaries between the units and other stratigraphic models. Using, when possible, the data of physical and other methods (magnetostratigraphy, fission-track dating) increases the reliability of these models. The correlation of zones and beds distinguished on the basis of diatoms, mollusks, foraminifers, and other fossils can be exemplified by the scheme elaborated for the Eocene–Pliocene sequence of the Shmidt Peninsula (Gladenkov *et al.*, 1999). The scheme (see table) shows position of zones and beds of different rank and their correlation with regional stages. The latter should ideally have the versatile paleontological characteristics. This is what the geological practice should tend to.

The Correlation Scheme. The Paleogene and Neogene correlation scheme of Sakhalin was published in 1998 (*Resheniya...*, 1998). It comprises 17 columns corresponding to different regions of northern (9 columns), central–southeastern (2 columns), and southwestern (4 columns) Sakhalin and to the Kurile Islands (2 columns). The scheme presented here includes only 10 columns, but these give a good idea on the correlation of formations from different regions of the island

Cenozoic stratigraphic units (beds, zones, horizons) of the Shmidt Peninsula

Age		Formations	Mollusks: beds	Mollusks: horizons	Foraminifers	Diatoms	Radiolarians	Palynozones and assemblages	Regional stages
Pliocene		Pomyr	17	X	12	*N. koizumii*	–	IX	Pomyr
			16			*N. koizumii – N. kamtshatica*			
		Matituki	15	IX	11	–	–		Upper Nutovo
Miocene	Upper	Mayamraf	14	VIII	10	*N. kamtschatica*	–	VIII	
		Vengeri	13	VII	9		–		Lower Nutovo
			12				*L. nipponicum*		
		Kaskad			8	*T. schraderi*		VII	Okobykai
			11	VI		–	–		
	Middle	Pil'vo	10		7	–		VI	
					6	*D. hyalina*	*E. inflatum*	V	
						D. lauta	*D. sakaii*	IV	Dagi, Uina
	Lower	Tumi	9	V	5	*T. fraga*			
			8	IV	4	–	–	III	
			7						
Oligocene		Machigar	6	III	3	*T. praefraga*			Kholmsk, Arakai
			5			*R. vigilans*			
			4	II	2	–	–		
			3						
			2						
Eocene			1	I	1				

(Fig. 2). It should be pointed out that the given scheme, compiled by L.S. Margulis, B.A. Salnikov and me, differs from that issued in 1998, because ages of particular formations and their ranges are corrected on the basis of new data that appeared during the most recent years. As mentioned above, the scheme is based first of all on the most important key sections of Sakhalin, the majority of which was studied by the group mentioned above.

As it follows from the scheme, some of the horizons are reliably traceable throughout the study region despite some debatable moments of the correlation of individual formations in different regions. Moreover, it is clear that the oldest Snezhinka Horizon remains to be a thing in itself. The large time range and a probable hiatus within the horizon make us to regard it as a stratigraphic unit with precaution. It is not improbable that the Paleocene–lower Eocene strata of Sakhalin will be subdivided into several horizons.

The Sakhalin horizon can be correlated rather accurately with regional stages of the western Kamchatka thus depicting the general tendency and stage-by-stage geologic history of the Sea of Okhotsk region.

SOME PECULIAR FEATURES OF PAST BIOTIC EVENTS

In the Paleogene and Neogene, Sakhalin was a part of the ocean–continent transitional zone that included a very complex system of sea basins, the evolution of which was associated with active tectonic and sedimentation processes. A combination of successive and discrete geological events of the global and regional significance was that very background, against which the flora and fauna of the zone evolved. Specific paleogeographic and paleoclimatic conditions and their fluctuations with time and in space controlled the evolution of shelf biota assemblages. In spite of complex environments and certain drastic situations in the past, the ecosystem remained to be rather stable. The composition and structure of biotic communities exhibit gradual changes, although the general trend was violated and evolution accelerated at certain moments.

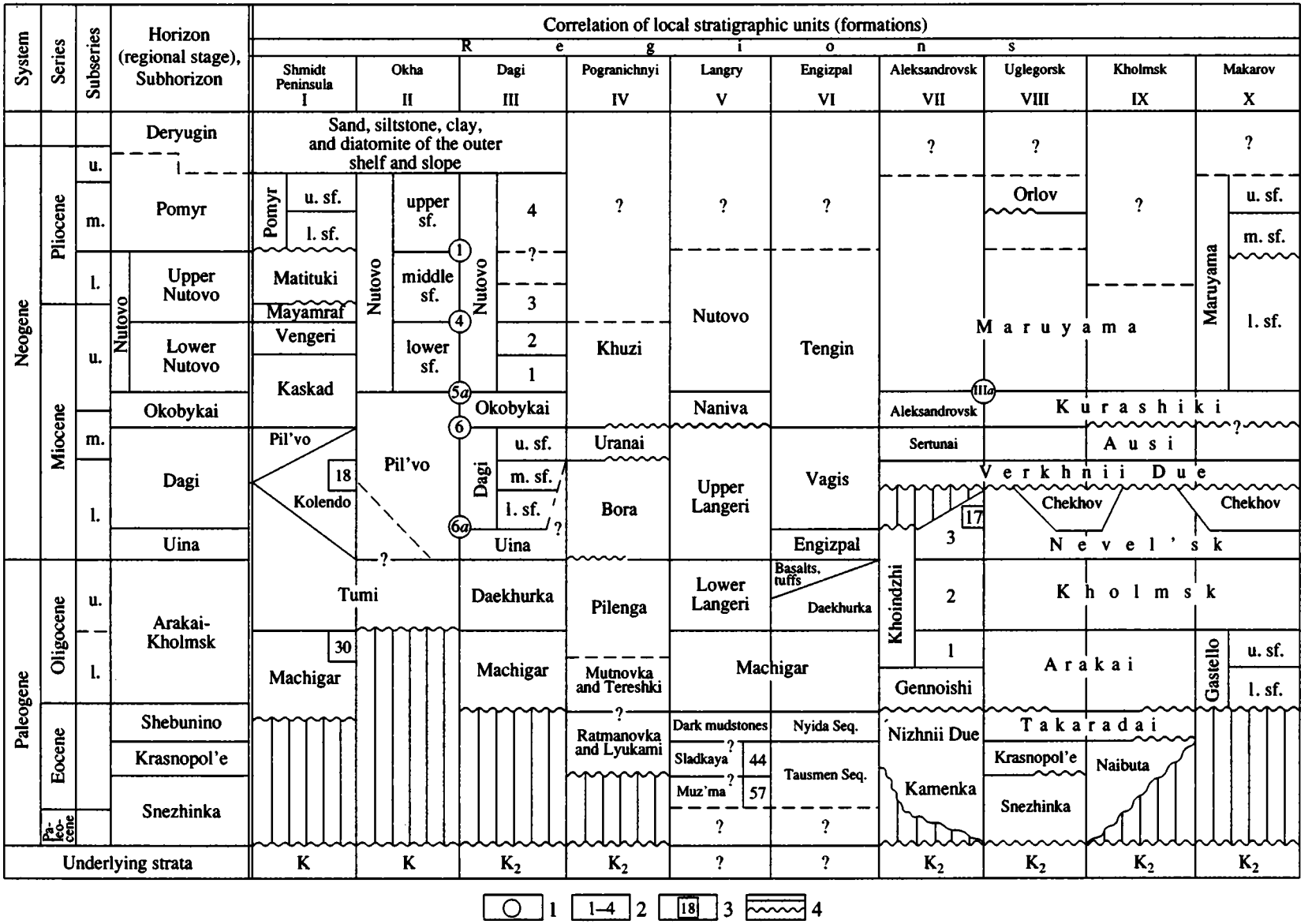

Fig. 2. The Paleogene–Neogene correlation scheme for Sakhalin: (1) seismostratigraphic reference horizons; (2) sequences; (3) radiometric dates (Ma); (4) conformable or unconformable bedding with washout. Subseries and subformations (sf.): (l) lower, (m) middle, (u) upper.

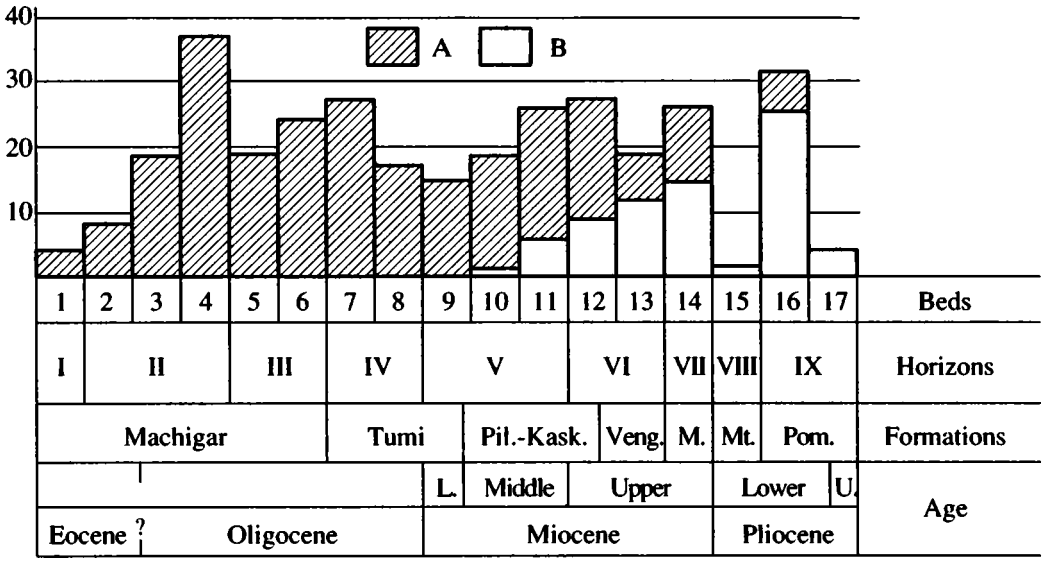

Fig. 3. Secular abundance variation of extinct and recent molluscan forms in beds (1–17) and horizons (I–IX) of the Shmidt Peninsula section, northern Sakhalin (Gladenkov *et al.*, 1999). Species: extinct (A) and recent (B). Formations: (Pil.–Kask.) Pil'vo, Kaskad, (Veng.) Vengeri, (M.) Mayamraf, (Mt.) Matituki, (Pom.) Pomyr. Subseries: (l) lower, (u) upper.

The analysis of benthic biota reveals the stage-by-stage evolution of the ecosystem recorded in successive fossil assemblages. I mean not mere changes of the latter throughout the section, which are related to different reasons, such as the basin depth variations, or shoreline displacements and migration of taxa in response to paleoclimatic fluctuations. What I mean is the succession of assemblages reflecting the general course of the benthic community evolution within the Sakhalin ecosystem (tidal–subtidal–bathyal) that was a part of the Sea of Okhotsk and North Japanese systems during the last 65 m.y. Four to five large stages (with several substages) can be distinguished in the community evolution, and they correspond well to the stages established in the Kamchatka.

Data on mollusks (as well as on benthic foraminifers) can be used to illustrate the biota evolution. Mollusks representing one of the dominant Cenozoic groups can be considered as index fossils. Within the extratropical zone of the northern hemisphere, mollusks prevailed in most of bottom biocoenoses of the Cenozoic basins producing the greater part of their biomass. The initial (Paleocene–earliest Eocene) stage remains yet unclear in the Cenozoic history of Sakhalin because of lack of corresponding data (see below). Subsequent middle–late Eocene, Oligocene, Miocene–early Pliocene, and late Pliocene stages are characterized by changes in generic and, especially, in species composition of assemblages (the renewal degree within horizons is up 30–50%). These stages also reveal variations in abundance of extinct (and currently living) forms (abundance rate of recent forms varied from 10 to 50% in the Miocene, and from 60 to 95–96% in the Pliocene). Figure 3 shows changes in the number of extinct forms throughout the section of the Shmidt Peninsula.

The stages, which were rather long and ranged from 11 to 17 m.y. except for the Late Pleistocene stage that is actually the beginning of a new stage, can be subdivided into substages (10–11 in total) each manifested themselves in assemblages of the horizons. The time ranges of substages were usually as long as 2–5 m.y. (an arbitrary assessment). The stages and substages depict the general tendency of the taxonomic diversity increase in boreal assemblages, peculiar morphological changes in shells of some molluscan phyletic lines, the growing abundance of mollusks in biocoenoses, specific rearrangements in the structure of the latter, etc.

Thus, it is safe to state that stages of biota evolution were of various scale and significance. Stages of the lowest order correspond to beds with fauna, the succession of which in sections reflects variations in paleogeographic and paleoecological conditions. The changing paleocoenoses of the beds are devoid of evolutionary changes of taxa and closely approach to stages of paleosuccessions. These are likely equivalents of ecozones. Their time ranges are assessed to be greater than hundreds of thousand years.

Stages of the second order correspond to local zones and horizons. Their succession reflects evolutionary transformations of taxons and paleocoenoses, although each horizon, as indicated above, depicts certain paleogeographic and paleoclimatic parameters of a given region as well. Their time ranges span several million years.

Stages of the third order characterize large events in the paleobiota history, when fauna experienced remarkable changes not only in Sakhalin, but also in the Sea of

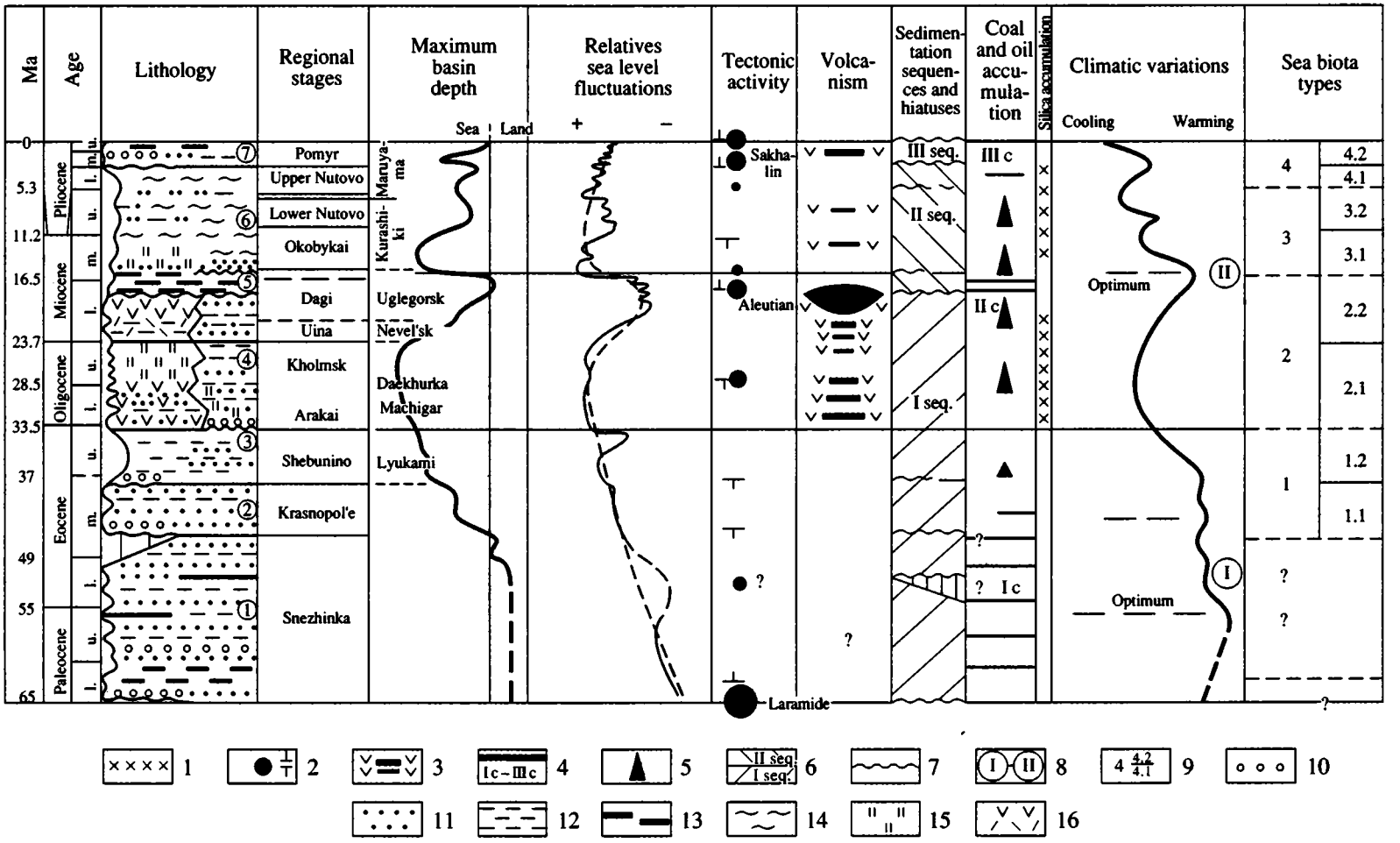

Fig. 4. Correlation of Cenozoic geological events in Sakhalin: (1) Silica accumulation stages; (2) phases of tectonic activation; (3) volcanic activity; (4) major epochs of coal accumulation; (5) oil and gas accumulation; (6) major sedimentation cycles; (7) washouts, hiatuses; (8) major climatic optima; (9) benthic biota types; (10) conglomerate; (11) sandstone; (12) siltstone, mudstone; (13) coal; (14) diatomite; (15) opoka, chert; (16) volcanic rocks and tuffs.

Okhotsk and North Japanese provinces, and even in the whole North Pacific Boreal Realm. For instance, the Oligocene–early Miocene stage characterizes the formation period of biota that developed later in the Neogene against the background of generation of the cryo- and psychrosphere of the Earth. This event resulted in a substantial rearrangement of biocoenoses and their structure in the shelf and marginal seas of all oceans. The stages (1–4) and substages (1.1–1.2; 3.1–3.2; 4.1–4.2), which are marked by assemblages of corresponding horizons, are shown in Fig. 4.

When characterizing major stages of the Cenozoic history of the paleobiota, one should bear in mind an important circumstance. Marine deposits are usually thought to form in Sakhalin during the (middle?) Eocene according to the representative sea fauna of the Krasnopol'e Horizon (*Ostrea, Cardium*, and other assemblages) regarded as the oldest in the island. In fact, this fauna begins the "paleosuccession" of Eocene–Pliocene biotic associations of the region. However, we should remember as well an interesting information about the oldest flora and fauna from the boundary Sinegorka Beds between the Cretaceous and Paleogene (Kalishevich *et al.*, 1981). The beds known in some Sakhalin regions (first of all, in the Naiba, Sinegorka, Manun, and other river basins) to rest (probably conformably) on the Upper Cretaceous marine deposits (usually on the Krasnoyarka Formation) and underlie conformable continental strata of the Paleocene "Nizhnii Due" Formation (the Snezhinka Horizon). They are composed of siltstones with interlayers of sandstone and mudstone (150–230 m), in which carbonate concretions are found.

In this case, molluscan assemblages include bivalves (35 species) and gastropods (13 species). According to Kalishevich, 78% of the assemblages are represented by recent genera. The lower part of the beds yielded representatives of specific Cretaceous genera (*Pseudaphrodina, Trigonocallista, Grammotodon*). Bivalves of Ctenodontida group dominate (six *Multidentata* and five *Truncacila* forms). Genera *Leionucula, Malletia, Modiolus, Glycymeris, Barbatia, Delectopecten, Limatala, Astarte, Thyasira*, and others are also present. The encountered gastropod genera are *Siphonalia, Psephaea, Cylichna.*

The assemblage differs from the Krasnopol'e fauna of the Eocene, and, hence, it was separated from the latter in time. It is hard to tell whether both assemblages are successive or not. The Sinegorka assemblage is likely to correspond to a special evolutionary stage of biota in the early Paleogene basins.

Unfortunately, the fauna and flora of the Sinegorka beds have not been studied in detail for about 20 years, though they are of great scientific and practical interest. A new study of the Sinegorka Horizon would contribute significantly to understanding of geological and paleontological problems of the Paleocene in the Far East. Today, new comparative material on the Paleocene of western Kamchatka has been published in a special monograph (Gladenkov *et al.*, 1997). The generalization of the material would elucidate the Paleocene stage of the biota evolution in the Sea of Okhotsk more reliably than it was done before. Lower Paleocene strata of western Kamchatka also yield fauna with individual representatives of genera passing to the Paleocene from Cretaceous beds (*Grammotodon* and others). Nevertheless, the lower contact with the Cretaceous deposits remains to be explored, as the beds transitional from the Cretaceous to the Paleogene and similar to the Sinegorka beds are unknown so far there.

Considering peculiar features of biocoenoses and their molluscan groups, I should mention as well that their distribution in sections depends on the vertical and trophic zoning. The tidal-upper subtidal molluscan communities appear to be most abundant and diverse. They are found to be variable and discontinuously distributed. By contrast, assemblages of the lower subtidal and bathyal zones are usually less diverse, but they are well traceable for large distances along shelf zones. The distribution of biocoenoses could be affected by the shelf morphology: a zonal distribution of biocoenoses is typical of steep shelves, where they regularly give way to each other with depth, whereas mosaic and spotted distribution patterns are characteristic of gentle shelves. All the aforesaid illustrate advantages and shortcomings of applying the paleontological material to stratigraphic purposes and paleogeographic reconstructions. In the case of extremely rapid accumulation of sediments (for instance, by the avalanche sedimentation in delta systems), biocoenoses become impoverished and many regular features of their "common" distribution disappear.

When analyzing paleocoenoses, one should take into account peculiarities of the trophic zonation, because it controls types of molluscan communities. This zonation reflects the adaptation of organisms from different taxonomic groups to a similar way of life and, first of all, to a similar feeding mode. Four trophic zones (groupings) are distinguished from a shore seawards: (1) sessile sestonophages, (2) vagile and sedentary sestonophages, (3) detritophages, (4) mud eaters. Each zone is characterized by a certain combination of different conditions providing certain nutrition type for organisms (food, water circulation, suspensions, abundance of organic matter at the bottom surface and in sediments, etc.). The first zone corresponds to the depth of 0–50 m, the second reaches the depth level of 100 m, the third and fourth occupy the depth interval of 100–200 and 300–500 m, respectively. The deciphering of biocoenoses in this context (calculation of taxons belonging to a certain trophic series, distinguishing of zones occupied by trophic groupings in space, etc.) allows us to reveal the history of shallow-water basins and successions of paleocoenoses originated due to different reasons (Zhidkova *et al.*, 1974). The analysis of Paleogene and Neogene paleocoenoses suggests that the trophic zones often changed synchronously in some

regions of Sakhalin. Although, it is a common case, when the shallow-water biocoenoses (with sessile sestonophages), for instance, those enriched in thermophilic forms from the Ausi Formation, prevailed in some regions, whereas the relatively deep-water groupings (detritophages), e.g., from the Pil'vo Formation, which look like cryophilic because of dwelling in deep-sea environments, dominated in other areas.

GEOLOGICAL EVENTS AND THEIR CORRELATION

There are two groups of geological events. Some of them were relatively short-term (volcanic eruptions, an abrupt sea level change, the appearance or disappearance of some groups of paleobiota, changes in the sea water temperature, etc.). These events are usually recorded in sedimentary sequences as certain markers ("protocols"), such as tuff interlayers, transgression and regression signs, changes in biotic communities, etc. It is the recognition of short-term (spasmodic) events within a large area that allows a correlation of sections of different structural–facies zones and ecosystems. Investigations of this kind represent subject of the event stratigraphy (Gladenkov, 1995).

Geological events include also some large-scale stages in the history of a given region, which are recognized on the basis of structural, depositional, paleogeographic, and other environmental changes. This general evolution by stages represents the background for development of individual structural–facies zones, and outlines regularities (or peculiarities) of the geological history of the region as a whole.

Not dwelling on this issue further, I would like to focus attention only on some background events, which played a significant role in the Cenozoic history of Sakhalin (Fig. 3). Important in this case are the following characteristics: general peculiarities in lithology of Paleogene and Neogene sections (with due regard to hiatuses and unconformities); depth changes in paleobasins and secular sea level fluctuations; periods of tectonic activation and volcanism; paleoclimatic events; particular epochs of coal and oil-and-gas accumulation; and peculiar features of biota evolution in paleobasins.

In general, the Paleogene–Neogene sedimentation in Sakhalin was of a cyclic character. The Cenozoic sedimentation history is subdivided into two major transgression–regression cycles. The third, Pliocene–Quaternary cycle is still in progress (Margulis, 1975). Each cycle begins and ends with continental coal-bearing deposits, whereas their middle parts are composed of marine sediments. The cycles are separated by the lower–middle Miocene coal-bearing formations, which mark the so-called Aleutian orogenic phase. Inside continental sequences, the boundary is unclear, but it is well defined in uplifts, where the middle Miocene marine deposits transgressively overlap Oligocene–lower Miocene strata with angular unconformity. In northern areas of Sakhalin, contrary to the southern ones, marine deposits form a single transgression–regression cycle almost uninterrupted by terrestrial sedimentation.

Consequently, accumulation of the Cenozoic marine sequences commenced in the Eocene (probably middle Eocene) time, and this moment signifies transition from continental to marine environments that was related to a large transgression in the Sea of Okhotsk region. The relative deepening of sea basins occurred in the Shebunino and, especially, the Kholmsk times (i.e., at the end of the Eocene and in the terminal Oligocene). The subsequent regression gave way to the accumulation of the Verkhnii Due carboniferous formation (the late Dagi time of the early Miocene) replaced afterward by a transgressive series. The second transgressive cycle culminated within the middle Miocene (the Okobykai time). An appreciable sea level rise also occurred at the beginning of the Pliocene (the middle Nutovo time). The Pomyr time of the late Pliocene corresponded to the regression period, except for a short phase of the marine regime at the beginning of it.

The cyclic evolution of sea basins naturally resulted in appearance hiatuses, erosion surfaces, and coal measures in sections. The most significant hiatuses of the regional scale are recorded at the basal levels of the Cenozoic (at the boundary with the Cretaceous), Verkhnii Due Formation, and Pomyr Horizon (I do not consider here the presumable hiatus within the Snezhinka Horizon). Hiatuses of the local scale are detected at the basal levels of the Krasnopol'e (?), Shebunino, and Arakai horizons, and also below the Kurashiki and Matituki formations.

Coal-bearing formations of Sakhalin accumulated in the lower parts of sedimentation cycles (the Paleocene–lower Eocene, latest early Miocene, and late Pliocene). The continental deposits and manifestations of coal are much less common in northern than in southern Sakhalin.

On the other hand, two phases of intense silica accumulation (beds of deep-sea siliceous deposits) are recognized in the marine sedimentation history. The first phase was in the Oligocene and early Miocene, whereas the second one corresponded to the middle–late Miocene. They both showed up in other region of the Sea of Okhotsk province as well (e.g., in western Kamchatka and Japan). The increased productivity of diatoms was responsible for appearance of abundant siliceous material during those phases.

It should be pointed out that the oil parental rocks mainly accumulated in the Oligocene and Miocene, when relatively deep-sea sequences were deposited in many regions of Sakhalin. The regional cover of oil-bearing strata and petroleum reservoirs of various types (from fissured siliceous rocks to inequigranular sandstone) formed at the same time. On the other hand, the activity of the ancient Amur River system in the Middle

and late Miocene created various reservoirs and traps for the organic matter.

A significant role in the formation and evolution of sedimentary basins of Sakhalin belonged to tectonic movements, which affected sedimentation and structural peculiarities in the island.

I.B. Pleshakov and his colleagues studied for a long time the Cenozoic folding and tectonic activation phases in Sakhalin (*Geologiya...*, 1970). They distinguished the Laramide phase at the boundary between the Cretaceous and Paleogene, the Aleutian phase of the pre-middle Miocene time, and the Sakhalin phase of the Pliocene–Quaternary. It is suggested that tectonic movements took place as well in the early (?) Eocene, Oligocene, and middle Miocene, when they changed to a certain extent the paleogeographic environments. Tectonic events governed many other geological phenomena, especially, their periodicity and general trend.

The definite cyclic nature of volcanic processes synchronous to tectonic and sedimentation events is inferred. Within the island, this is evident for the formation period of the Oligocene volcanic level in the Arakai Formation. Volcanism was still in progress in the Kholmsk and Nevel'sk time (Oligocene–early Miocene), when a great mass of tuffs accumulated. The volcanic activity culminated in the Chekhov time (the terminal early Miocene), when thick volcanic sequences accumulated in the southern Sakhalin during the Aleutian orogenic phase. Traces of the Miocene to Pliocene volcanic activity are recognized as well in other sections of Sakhalin (tuff beds of the Okobykai, Nutovo, and Pomyr times), but they are of minor significance. It should be emphasized that the Oligocene–early Miocene volcanism marks a substantial structural rearrangement in the northwestern part of the Pacific province. It is likely that the Japan–Sea of Okhotsk region had acquired features typical of the recent active margin by the end of that period.

Noteworthy also is the relationship between the relative seal level fluctuation and the deepening of sea basins adjacent to Sakhalin. Judging from curves plotted on the bases of seismostratigraphic data, one may sometimes infer a definite synchronism between these processes (the sea level rise leads to widening of deep-water sedimentation zones, which become narrower in the periods of sea level drop). In the Oligocene and Miocene, the maximum basin depth, usually corresponding to accumulation sites of siliceous–clayey sediments bearing the relatively deep-water biotic assemblages, correlates well with phases of the relative sea level rise and with descending tectonic movements. The regression periods and land uplifts were connected in time with main orogenic phases (for instance, with those of the early Miocene and Pliocene).

In general, sea level changes in Sakhalin basins probably reflect not only tectonic events but, in some cases, the global eustatic fluctuations as well. Influence of the latter presumably was significant for the transgression in Sakhalin about 15–16 and 10.5 Ma ago, or for regressions about 5 and 3.5 Ma ago. Some of these events are correlative with those recorded in Kamchatka. Unfortunately, sequences and parasequences, which may be correlated with global eustatic events, are identified reliably on the basis of seismic data only in the upper part of the section, mainly in the Pliocene and, partially, in the Miocene interval.

Along with various events recorded in Sakhalin, cyclic paleoclimatic changes and corresponding transformations of flora and fauna are also revealed in the Sakhalin sections. The Paleocene and initial Eocene are regarded as the warmest time intervals (Optimum I of the Cenozoic in Sakhalin). In general, the Eocene climate was warm except for the minor cooling at the end. The Oligocene corresponds to the coldest Paleogene epoch. The subsequent relative warming began at the end of the early Miocene and culminated at the beginning of the middle Miocene (Optimum II). Despite the lower paleotemperature parameters as compared to the Paleocene–Eocene warming peak, this optimum is recorded in many regions around the Sea of Okhotsk and in the whole North Pacific region. Climatic fluctuations in subsequent periods were smaller in scale (warm peaks are recorded, for instance, at the beginning of the late Miocene, in the middle Pliocene, etc.) and developed against the background of progressing cooling that approached its maximum in the Quaternary.

Clearly, paleoclimatic fluctuations controlled migration of paleobiota: subtropical assemblages with *Dosinia* and other thermophilic representatives appeared in Sakhalin basins during the Ausi warming phase of the Miocene. In contrast, cryophilic taxa (in particular, *Astarta* species) migrated in the study region during the Pliocene cooling.

Warming phases correlate in time with epochs of the relative sea level rise, though it is difficult to prove this for minor fluctuations. Quite different tendencies were characteristic of the Eocene and Oligocene. For instance, the Oligocene cooling coincided with the relative seal level rise.

Processes of paleobiota evolution come to light against the background of abiotic events of the last 65 m.y. In this case, I mean not the ecological succession of paleocoenoses related, in particular, to the basin depth changes, but the evolutionary transformations and reorganization of communities within the large ecosystem of the Boreal Realm. I already mentioned above the major cycles in the evolution of benthic communities, i.e., four stages (without the Sinegorka one) and eight substages, which are inferable from changes in paleontological assemblages. In some cases, the synchronism of stages with abiotic events is somehow detectable (for instance, the Oligocene stage 2 coincides in time with the relative sea level rise and cooling), but the cause-and-effect relations between these

events is far from being clear, and with respect to evolutionary substages it has not been established at all. We really need a deeper insight into this problem.

In conclusion, I may state that placing different geological events on the time scale we can understand where the events coincide and when such a coincidence is undetectable. Moreover, the revealed synchronism of some events does not mean always their cause-and-effect relations. The latter is a subject of specialized analysis.

On the basis of manifestation (and partial coincidence) of particular events, the following turning points are established in the geological history of Sakhalin: the middle Eocene (the onset of transgression and climatic warming), the Oligocene (deepening of the basins, volcanic activity, silica accumulation, formation of the Neogene fauna core, and cooling phase), the early Miocene (tectogenesis, intense volcanism, regression, and coal accumulation), the middle Miocene (new warming phase and deepening of the basins), the Pliocene (tectogenesis, regression, coal accumulation, formation of the of the modern fauna core, and warming–cooling episodes).

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Reviewer L.A. Nevesskaya

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Stratigraphic and Geographic Distribution of Late Eocene Foraminifers in the Northern Periphery of the Tethys

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Abstract—Foraminiferal assemblages from the Belaya Glina basin, which extended from the Black Sea to the Aral–Turgai region and eastern Central Asia, and from the inner seas of Transcaucasia are analyzed. The assemblages mainly consist of species migrated from the Mediterranean after the fauna extinction at the pre-Priabonian time. The most representative assemblages are established in the Crimean–North Caucasian region, where the *Planulina costata* Zone subdivided into the *Brotzenella taurica* and *Bolivina antegressa* subzones is distinguished on the basis of benthic foraminifers as overlain by the *Almaena taurica* or *Nonion curviseptum* beds. The climate- and facies-dependent variability of fauna is detected to the north and east of the region, but the guide species of the *P. costata* Zone were found everywhere. These species are used to correlate the local biostratigraphic subdivisions to each other and with the units of the standard stratigraphic scale. Changes in the foraminiferal assemblages from the Priabonian boundary beds of different-facies sections are considered. The Belaya Glina basin corresponded to the northern province of the Mediterranean paleobiogeographic realm, where the Tajik and East Transcaucasia subprovinces are recognized.

Key words: foraminifers, upper Eocene, distribution, assemblage variability, paleobiogeography, the Tethyan basin.

INTRODUCTION

In the former USSR territory, the upper Eocene (Priabonian) marine deposits conformably lie on marls of the Kuma Horizon of the Bartonian Stage but in some areas they unconformably overlie older (even Upper Cretaceous) deposits. The upper Eocene deposits show clear lithological distinctions. They are usually represented by clay limestones, marls, and carbonate clays, which are light-colored, nearly white in the southeastern North Caucasus and rose-colored in Turkmenistan and western Kazakhstan. Facies of shallower settings are carbonate and silty clays in the Ciscaucasia and Russian Plain, or clayey and silty deposits in the northern Aral region, northern Ustyurt and southern Turgai. In the Transcaucasia, sections are composed of marls and carbonate clays. In the southern West and Central Europe, outside the former USSR, the Priabonian sections consist of carbonate sediments (including reef facies).

The upper Eocene deposits are attributed to the Belaya Glina Horizon (regional stage) in the Crimea–Caucasus region and Turkmenistan, to the upper part of the Solon Formation in the Volga region and the Ergeni Upland, to the Adaevskaya Formation in the Mangyshlak Peninsula, and to the Chegan Formation in northern Kazakhstan. The deposits persistently yield the diverse assemblages of benthic foraminifers, which enable the direct correlation of remote sections devoid of plankton. The fauna composition was controlled by the latitudinal–climatic zoning and by the facies confinement that affected structures of paleobiocoenoses.

Ambiguity in species identifications is responsible for a certain loss of information and makes it difficult to compare the available lists of fossils even from the neighboring parts of the paleobasin. The data published do not always adequately characterize composition of the assemblages and cannot serve as a reliable basis for stratigraphic and paleobiogeographic conclusions. In this work, data on regional fauna are obtained from paleontological collections with due account for the additional published information. Data on faunas from foreign countries are quoted from respective publications.

PALEONTOLOGICAL CHARACTERISTICS

Assemblages of planktonic and benthic foraminifers from the northern periphery of the Tethyan basin mostly consists of species migrated from the Mediterranean. Accordingly, areas adjacent to the Mediterranean region (Armenia, Azerbaijan) are characterized by warm-water genera unknown in the northern regions. The core portion that can be called "the Belaya Glina fauna" was formed in the Crimea–Caucasus region and then dispersed widely. Toward the north and northeast, the assemblages show decreasing taxonomic diversity due to latitudinal–climatic factor and composition of sediments depleted in carbonate. Flysch deposits of the Carpathians in Ukraine mainly contain agglutinating

foraminifers, but carbonate facies also yield here their secreting species characteristic of the Crimean–Caucasian fauna.

Planktonic Foraminifers

The impoverished assemblages of planktonic foraminifers from the region under consideration are close in composition to the Mediterranean fauna. This became evident, when *Hantkenina suprasuturalis* Bronn. and *H. dumblei* Cushm. et Weinz. earlier unknown in the Belaya Glina assemblages were discovered. In the Armenian sections, where fauna is most similar to the Mediterranean one, V.A. Krasheninnikov distinguished the *Globigerapsis semiinvoluta, Globorotalia cocoaensis*, and *G. centralis* zones (*Geologicheskie...*, 1996). In the northern sections, where two first species were not found, the *Globigerapsis tropicalis* s.l. Zone frequently subdivided into two parts was established. This zone shows everywhere the decreasing abundance of planktonic forms upward in the section, where "larger globigerins" are replaced by "smaller ones," whose prosperity time was in the early Oligocene.

In the relatively deep-water deposits of northern Caucasus (the Kuban' and Kheu river basins), two parts of the *Globigerapsis tropicalis* s.l. Zone were considered as independent zonal units by Krasheninnikov and as the *G. tropicalis–Globigerina corpulenta* and *Turborotalia centralis* subzones by myself (Bugrova *et al.*, 1988). In the Crimea, Mangyshlak, and some areas of Turkmenistan, these subdivisions can be only ranked as "beds" because of the indistinct boundary between them. In the shallow-water sections of the Russian Platform, Scythian plate, and platform part of Turkmenistan, the distribution of *G. tropicalis* is confined to basal beds, and above, where this species disappears, subdivisions named after *Globigerina praebulloides* are distinguished (Nikitina, 1972; *Paleogen...*, 1975). The lower part of the *G. tropicalis* s.l. Zone contains *Globigerapsis tropicalis* Bann. et Blow, and *Globigerinoides rubriformis* Subb. in association with *Globigerina corpulenta* Subb., *Globigerina inflata* Orb., and *G. subtriloculinoides* Chal. ("larger globigerins"). In the upper part of the zone, the former three species disappear being replaced by *Turborotalia centralis* (Cushm. et Berm.), *Globigerina gortanii* Bors., *G. galavisi* Berm., *G. angiporoides* Hornibr., and also by "smaller" *Globigerina officinalis* Subb. *G. praebulloides* Blow, and *G. ouachitaensis* Howe et Wall., which associate in the Oligocene deposits with *Globigerina khadumica* N. Byk. and *Tenuitella liverowskayae* (N. Byk.).

Benthic Foraminifers

In some regions, benthic foraminifers were used to distinguish local subdivisions in the rank of "beds" or "zones." The subdivisions were locally united into regional (local) zones, e.g., in the southern Russian platform and Scythian plate (Nikitina, 1972), in Turkmenistan (Bugrova, 1979), and Tajikistan (Davidzon *et al.*, 1982). The studied collections of fauna from these subdivisions showed that they contain the same guide species, and this was used to distinguish the upper Eocene *Planulina costata* Zone in the southern part of the former USSR (Bugrova, 1988b, 1992). The zonal assemblage has persistent composition within a vast territory, the Transcaucasia included. Many of its guide species were initially described from sections of southern and central Europe.

In most areas, the assemblage includes *Cyclammina pseudocancellata* Chal. (?=*C. rotundidorsata* Hantk.), *Vulvulina pectinata* (Hantk.), *Cylindroclavulina terterensis* (Chal.), *Clavulinoides szaboi* (Hantk.), *Karreriella hantkeni* Cushm., *Robulus budensis* (Hantk.), *R. arcuatostriatus* (Hantk.), *R. limbosus* (Reuss), *Marginulina bohmi* (Reuss), *Marginulinopsis infracompressus* (Thalmann), *Turkmeniella kubynii* (Hantk.), *Alabamina almaensis* (Samoil.), *Eponides umbonatus* (Reuss), *Anomalina granosa* (Hantk.), *Anomalinoides affinis* (Hantk.), *A. alazanensis* (Nuttall), *Brotzenella taurica* (Samoil.), *Heterolepa dutemplei* (Orb.), *H. pygmea* (Hantk.), *Planulina costata* (Hantk.), *Bulimina truncana* (Guembel), *Uvigerina eocena* Guembel, *U. jacksonensis* Cushm., *Grammostomum nobilis* (Hantk.), *Globocassidulina globosa* (Hantk.), and some other species, all characterizing the entire zone.

In many regions, the zone is subdivided into two subzones. The lower *Brotzenella taurica* Subzone contains some species of the zonal assemblage in association with *Brotzenella taurica* (Samoil.), *Tritaxilina hantkeni* Cushm., and some *Robulus* forms characteristic of the subzone. The index species of the subzone was found in lower beds of the sections exposed in Crimea, northern Caucasus, Georgia, Armenia, Moldova, northern Black Sea region, eastern coast of the Sea of Azov, Stavropol' and Volgograd areas, Krasnovodskii Peninsula, western chinks of the Ustyurt and Mangyshlak, and also in the northern coast of the Aral Sea.

Diverse *Bolivina* forms are characteristic newcomers of the upper *Bolivina antegressa* Subzone, and this feature allowed Subbotina to term this unit originally as the *Bolivina* Zone. However, abundance of *Bolivina* forms and decreasing content of planktonic foraminifers turned out to be facies-dependent. Species that appear in the subzone are *Spiroplectammina azovensis* J. Nik., *Karreriella hantkeni* Cushm., *Robulus calcariformis* J. Schwemb., *Oridorsalis schreibersi* (Orb.), *Epistomina paleogenica* Mjatl., *Svratkina perlata* (Andreae), *Ceratobulimina intrusa* N. Byk. (?=*C. contraria* (Reuss), *Cassidulina laevigata* Orb., *Bolivina antegressa* Subb., *B. budensis* (Hantk.), *Grammostomum nobilis* (Hantk.), *Fursenkoina halkyardi* (Cushm.), and others. In general, the assemblage of the upper subzone dwelt in shallower waters. Species *Nonion curviseptum* Chal., *Almaena taurica* Samoil.,

Pseudogaudryina trigona (Ter-Grig.), *Robulus diaphanus* (N. Byk.), *Melonis dosularensis* (Chal.), *Cibicidoides kugultaensis* (Schutzk.), *C. jankulaensis* (Schutzk.), *Asterigerina falciloculari*s Subb., *Vsegeina nana* Bugr., *Angulogerina angulosa* (Will.), *Tergrigorjanzaella sectile* (Ter-Grig.), *Rectuvigerina lacera* (Subb.), *Cassidulinoides bodeni* Hagn., *Chilostomelloides caucasicus* Chal., and others appear in some regions near the top of the subzone. Some of them pass into the lower Oligocene beds. According to species distribution in the topmost part of the zone, the *Almaena taurica* Beds are distinguished in the Crimea and their analogues (*Nonion curviseptum* Beds) are recognized in the northern Caucasus, Transcaucasia, northern Kara-Bogas area, western chinks of Ustyurt, Mangyshlak, and in the southern coast of the Aral Sea. Shutskaya reported that index species of two last units are concurrent in Daghestan (the Rubaschai River section).

Outside the former USSR, species characteristic of the *P. costata* Zone were found in the Priabonian deposits of Poland (Pozaryska, 1978), Bavarian Alps (Hagn, 1956; Hagn and Lindenberg, 1966), and northern Italy (Braga *et al.*, 1975; Grunig, 1985). Many species of the zone are characteristic of the "*Clavulina szaboi* Beds" near Budapest (Hantken, 1875) and also are known in sections of Spain (Ruiz de Gaona and Colom, 1950). According to Krasheninnikov, they occur in the upper Eocene sections of Egypt and Syria. In the Priabonian stratotype area, subdivisions based on distribution of benthic species have not been established, but zones of planktonic foraminifers still bear some benthic forms (Braga *et al.*, 1975). Two lower zones yield *Marginulina bohmi, Robulus arcuatostriatus, Globocassidulina subglobosa, Uvigerina eocena, Grammostomum nobilis*, and other species of the *P. costata* Zone. In the upper *Globorotalia cunialensis* Zone, there appear nummulites, Miliolidae, *Asterigerina* and *Pararotalia* forms indicative of the basin shoaling like in the Crimea–Caucasus region. In northwestern Europe, the Priabonian corresponds to an upper part of the *Nummulites-Eponides* Subzone FH1 (Doopert, 1988) bearing fauna distinct from that of the *P. costata* Zone.

BOUNDARIES OF THE UPPER EOCENE

Boundaries of the Priabonian could be reliably defined, if phylogenetic lineages of species that evolved for a long time under stable conditions were known. However, conditions in marginal parts of the basin are unstable, and the boundaries should be recognized here on the basis of compositional changes in the assemblages.

The *lower* boundary is defined everywhere at the first occurrence level of plankton of the *Globigerapsis tropicalis* s.l. Zone and nannoplankton of the *Discoaster barbadiensis* Zone, which also marks the compositional renewal of benthic genera and families. In case of simultaneous faunal and lithological changes, this boundary is well distinguishable in both shallow- and deep-water sections.

So, in the Volga–Don interfluve (Borehole 14813, the Melikhov hydrosystem, information of 1951 by M.E. Zubkovich), there were recovered carbonate-free clays of the Kuma Horizon, which yield *Caucasina* sp., *Schenckiella gracilis* Bugr., and other arenaceous foraminifers in association with radiolarians and sponge spicules. In the overlying carbonate-bearing deposits, the late Eocene planktonic and benthic (71 species belonging to 51 genera) foraminifers appear simultaneously (Fig. 1). Species most characteristic among them are *Neoeponidella lucida* (Minak.), *Marginulina bohmi* (Reuss), *Planulina costata* (Hantk.), *Uvigerina jacksonensis* Cushm., *Bulimina truncana* Guemb., and *Grammostomum nobilis* (Hantk.). Arenaceous tests of *Spiroplectammina azovensis* J. Nik. and *Karreriella hantkeni* Cushm. represent a large share of the assemblage. *Brotzenella taurica* occurs in the lower part of the section, and *Bolivina antegressa* appears in the middle of the section together with species characteristic of the upper subzone.

In some places, forms of the new assemblage appear at the end of the middle Eocene, but this level may also correspond to distribution range of recurrent fauna in other places. The recurrence of older assemblages and similar facies hampers the boundary recognition. This phenomenon was recorded by Ter-Grigor'yants *et al.* (1968) in the Stavropol' area and by Kurgalimova and Beniamovskii in the northern Ustyurt and Caspian depression. In southeastern Turkmenistan (Timofeev *et al.*, 1972), carbonate deposits contain the assemblage of the *P. costata* Zone, whereas slightly calcareous clay interbeds predominantly enclose species of the underlying zone.

Bykova (1974) thoroughly studied the facies-dependent faunal changes across the boundary beds between the Shorym (middle Eocene) and Adaevskaya (upper Eocene) formations of the Bayurbas section (western chinks of the Ustyurt). The boundary interval yielded an assemblage of mixed composition, and Bykova distinguished the "*Bolivina asiatica–Astacolus asangulensis*" Beds with an older assemblage in the lower part of the upper Eocene. In the same region (the Kenderly area), light-colored marls of the upper Eocene include basal interbeds of dark-colored marls similar to those of the Kuma Formation and still containing rare tests of *Globigerina turcmenica.*

When we have "transitional beds" with mixed fauna, the boundary position can be defined ambiguously on the basis of different fossil groups. A good example is the Mangyshlak section (the Uzunbas Ravine of the Karagie depression, data of V.I. Zhelezko, 1988). Brown marls of the Shorym Formation bearing here fish scales and teeth (Fig. 2) also yield *Truncorotaloides rohri* Bronn. et Berm. and species of the *Globigerina turcmenica* Zone and nannoplankton of Zone NP17 in association with benthic for-

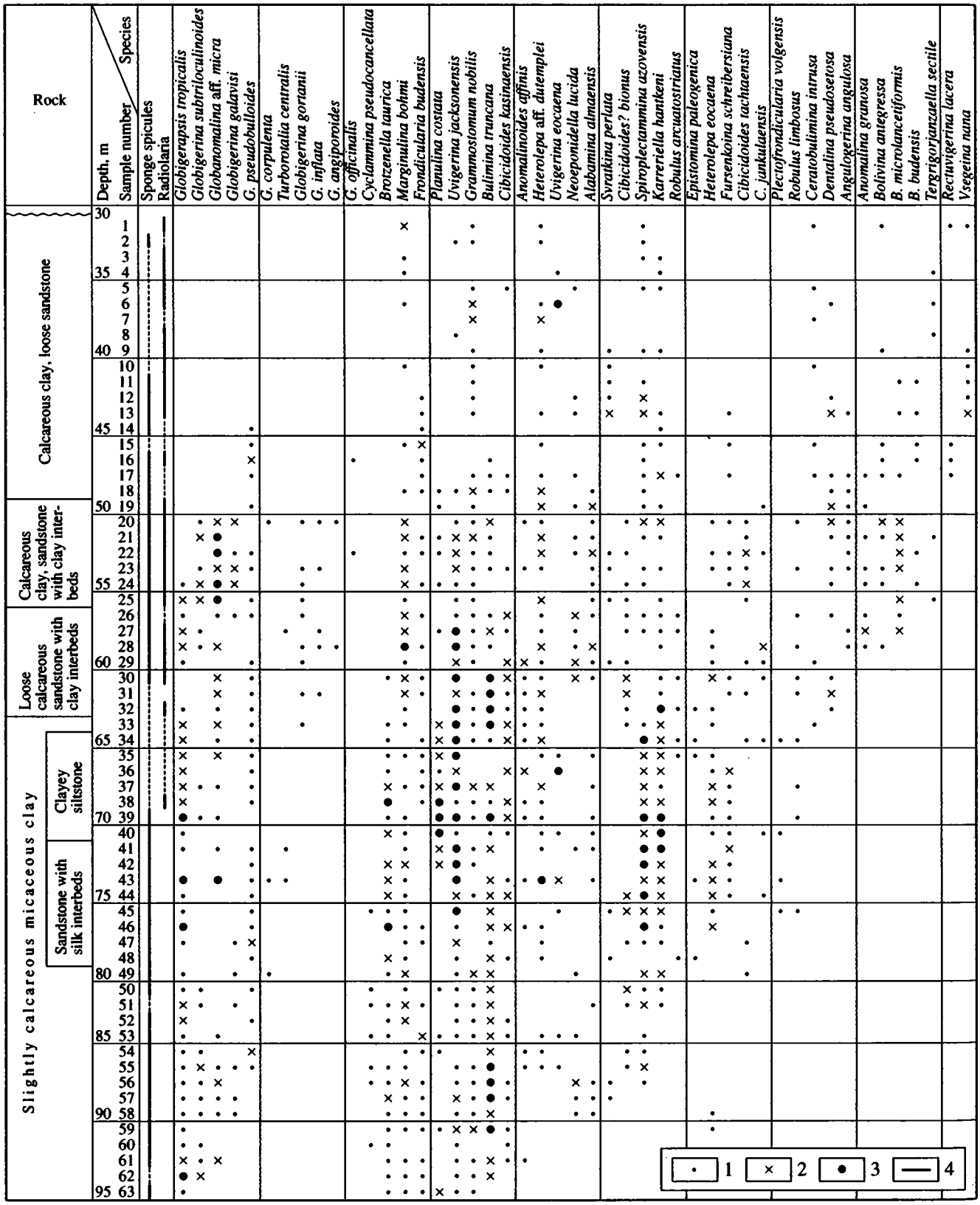

Fig. 1. Distribution of late Eocene foraminifers in the section of Borehole 14813 (the Melikhov hydrosystem). Abundance rates of foraminifers: (1) low; (2) medium; (3) high; (4) undetermined.

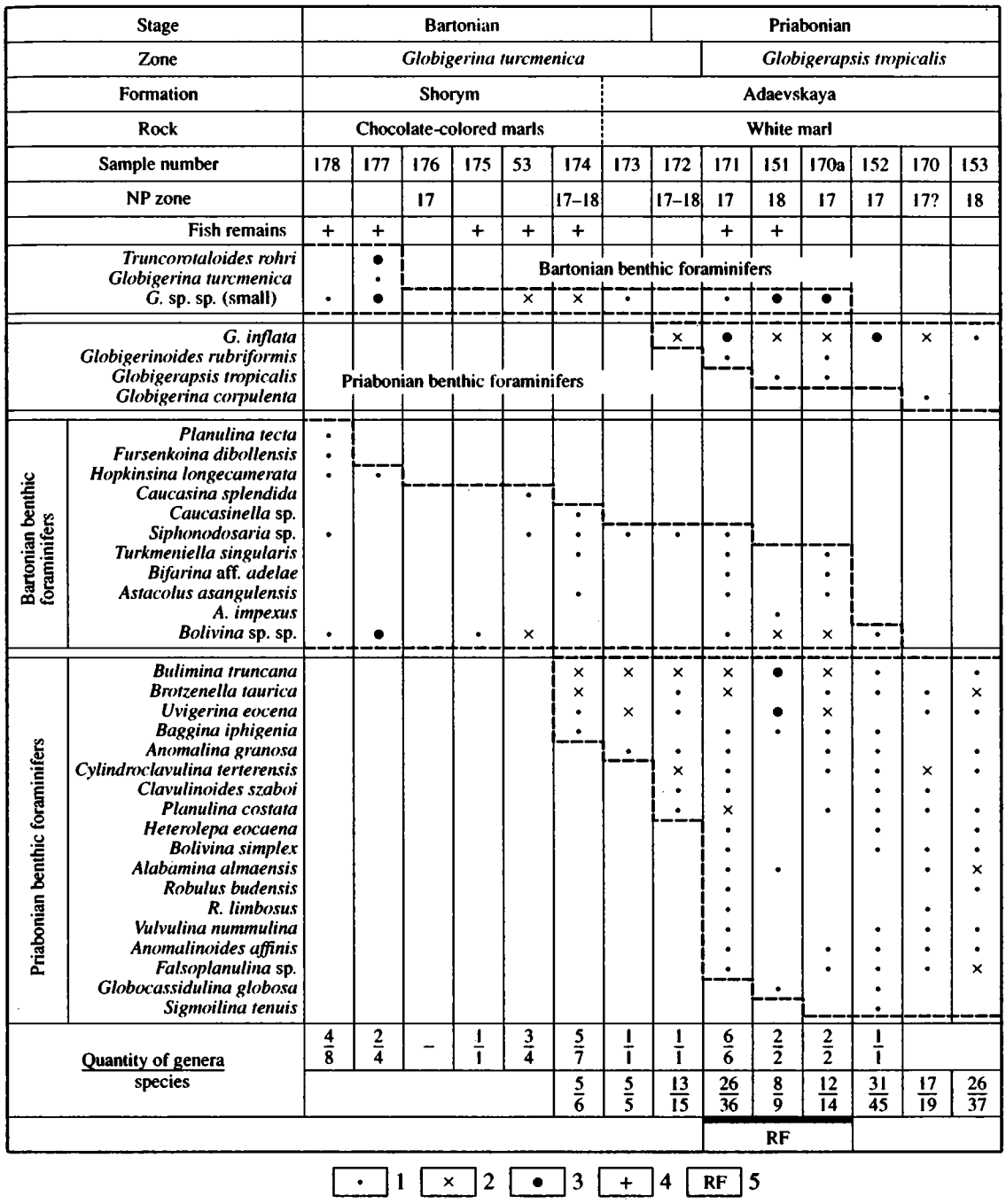

Fig. 2. Quantitative taxonomic composition and successions of foraminiferal assemblages in the Bartonian–Priabonian boundary beds (data of 1988 by V.I. Zhelezko characterizing the Uzunbass Ravine site in the Karagie depression of the Mangyshlak Peninsula). (1–3) abundance rates of foraminifers: (1) low; (2) medium; (3) high; (4) fish remains; (5) recurrent fauna.

aminifers of the *Caucasinella pseudoelongata* Zone. These zones represent subdivisions of the Bartonian Stage. Near the formation top (Sample 174), I.P. Tabachnikova recorded the first occurrence of the index species characterizing Zone NP18 and formally representing marker of the lower Priabonian boundary. Benthic foraminifers of the *P. costata* Zone appear at the same level. Their abundance increases upward from 46% (Sample 174) to 93% (Sample 172). In this section, the boundary is defined not by the first occurrence, but by the diversification level of zonal assemblage of the *P. costata* Zone. Slightly above, there appear planktonic forms of the *G. tropicalis* Zone. Such a divergence of boundaries marked by different fossil groups is unimportant in practical aspect, because the interval thickness is insignificant.

Benthic forms of the middle Eocene appear again in the basal dark-colored clay interbeds of the Adae-

vskaya Formation. Nannoplankton assemblages lacking here zonal species of the Zone NP18 are similar, in opinion of Tabachnikova, to assemblages of the underlying zone. Beginning from the level of 1.1 m above the base of the Adaevskaya Formation (Sample 171), no middle Eocene fauna was found. Thus, the repeated occurrence of foraminifers and nannoplankton species is related here to relapse of environments favorable for their development.

The mixed assemblage was also found in the southern Emba area (Borehole 448-Tengiz, interval of 109–105 m, data of S.N. Nikolaeva). At the depth of 109 m, gray clays contain abundant *Globigerina bulliformis, G. praebulloides*, and other planktonic species of the *G. turcmenica* s.l. Zone in association with rare *Globigerina inflata* and *Globigerina corpulenta* characteristic of the succeeding zone. The benthic group includes 11 Bartonian and two Priabonian species. Tabachnikova recognized there the NP17 Zone. Above (depth of 108.5 m), deposits are white. At this level, abundant smaller globigerins and *Globanomalina micra* (Cole) typical of the upper part of the *G. turcmenica* Zone are associated with the Priabonian species *Globigerina inflata* Subb. and *Globigerina corpulenta* Subb. Benthic group includes numerous "smaller bolivins" of the Kuma Horizon (*Bolivina serrata* Chal., *B. calatha* N. Byk. et Chram., *B. semiintegra* Chal., *B. dilatabilis* Chal.) together with *Planulina tecta* Rod., *Astacolus impexus* N. Byk., *A. asnagulensis* N. Byk., *Siphonina praelata* N. Byk., and others (11 species altogether). In addition, the level yielded 24 scarce species (almost 70% of the assemblage) of the *P. costata* Zone, *Brotzenella taurica* (Samoil.), *Planulina costata* (Hantk.), *Alabamina almaensis* (Samoil.), *Cylindroclavulina terterensis* (Chal.), and some other forms included. The higher level of 105 m yielded the benthic assemblage (25 species of 20 genera) of the *P. costata* Zone and *Globigerapsis tropicalis* appeared for the first time. In this section, the lower Priabonian boundary can be drawn at the depth of 108.5 m, where the Priabonian plankton is present, species of the *P. costata* Zone are dominant, and the first occurrence of nannoplankton of the NP18 Zone is registered. It seems that some calcareous Bartonian species were able to adapt sedimentation environments changing from terrigenous to carbonate and existed some time together with younger species.

In the northern Ustyurt chinks, where no plankton was found, the boundary is drawn at the level of compositional changes in benthic assemblages. At the base of the Chegan Formation, species of 20 genera of the *P. costata* Zone (Bronevoi *et al.*, 1967, Exposure 509, collection of N.N. Bryzzheva) replace the assemblage of arenaceous foraminifers *Popovia hirta* Suleim., *Bolivinopsis turgaica* Balakhm., *Gaudryinopsis superturkestanica* (N. Byk.), and others typical of the Kuma Horizon. In the southern Turgai (map sheet M-41-XXV, Borehole 51, data of N.I. Komarova), the species listed above are associated with *Cyclammina pseudocancellata* Chal. dominant in the late Eocene assemblage of synonymous "beds" (or "zone" in interpretation of I.M. Aizenshtat).

The *upper* boundary of the *P. costata* Zone coincides in the Northern Caucasus and Central Asia with the lithological transition from carbonates to slightly calcareous and carbonate-free clays of the Oligocene age. These regions isolated from the Tethys in the Oligocene time are characterized by several benthic assemblages changing in the lateral direction in their species and even generic composition (Pechenkina, 1971a; *Reshenie...*, 1970). The boundary is variably defined here. In the areas still connected with the Mediterranean basin (Armenia), the boundary is recognized on the basis of planktonic foraminifers. For instance, Krasheninnikov distinguished the *Globigerapsis semiinvoluta, Globorotalia cocoaensis*, and *Turborotalia centralis* zones of the Priabonian and the *Globigerina tapuriensis* Zone of the lower Oligocene in the Landzhar section. In contrast to the Crimean–Caucasian fauna, the highly diverse benthic assemblage from this section contains some warm-water genera. The fauna in question can be considered as the Tethyan one mixed with some elements of the Belaya Glina basin.

Benthic assemblages show gradual changes across the boundary. The Eocene carbonate deposits contain many species of the *P. costata* Zone associated with warm-water *Pseudoplanulina demesini* (Debourle et Sacal), *Halkyardia minima* (Lieb.), *Stomatorbina enodata* (Chal.), *Pararotalia octocamerata* (Chal.), *Chapmanina gassinensis* Silv., and *Caucasina aziderensis* Chal. Species typical of the Eocene deposits are *Vulvulina pectinata* (Hantk.), *Cylindroclavulina terterensis* (Chal.), *Plectofrondicularia striata* (Hantk.), *Hydromylina terreyi* (Coryell et Embich), *Robulus budensis* (Hantk.), *Turkmeniella kubinyii* (Hantk.), *Astacolus cymboides* (Orb.), *Marginulina subbulata* Hantk., *M. Bohmi* (Reuss), *M. eocenica* (Franke), *M. vasarhelyi* (Hantk.), *Lagena biarritzensis* Hagn., *Frondicularia tenuissima* (Hantk.), *Osangularia mexicana* (Cole), *Asterigerina falcilocularis* Subb., *Valvulineria palmarealensis* Katch., *Anomalina granosa* (Hantk.), *Planorbulina* aff. *mangyshlakensis* Vassil., *Uvigerina striata* Costa, *Rectuvigerina lacera* (Subb.), *Bolivina antegressa* Subb., *Cassidulina* aff. *havanensis* Cushm. et Berm., *Bulimina truncana* Guembel., and *Rectobolivina* sp.

Some species pass from different Eocene levels into the *G. tapuriensis* Zone. These are *Clavulinoides szaboi* (Hantk.), *C. jarvisi* (Cushm.), *Pseudogaudryina trigonia* (Ter-Grig.), *Karreriella siphonella* (Reuss), *K. disjuncta* (Cushm. et Jarv.), *Matanzia* sp., *Siphotextularia* sp., *Chrysalogonium longicostatum* Cushm. et Jarv., *Cancris erevanensis* (Saakjan), *Oridorsalis umbonatus* (Reuss), *Vsegeina nana* Bugr., *Siphonina reticulata* (Czjzek.), *Falsoplanulina* sp., *Cibicidina westi* (Howe), *Planulina costata* (Hantk.), *Anomalinoides alazanensis* (Nutt.), *Cibicidoides* aff. *jankulaen-*

sis (Schutzk.), *Heterolepa almaensis* (Schutzk.), *H. dutemplei praecincta* (Karrer), *Queraltina epistominoides* Marie, *Melonis dosularensis* (Chal.), *Turrilina alsatica* Andreae, *Bulimina alazanensis* Cushm., *Trifarina labrum* Subb., *Aragonia reticulata* (Hantk.), *Bolivina beyrichi* Reuss, *B. maragaensis* Chal., *B. prebinaensis costata* Chal., *B. saltakensis* Chal., *Loxostomoides millepunctatus* (Tutk.), *Cassidulina laevigata* Orb., and *Globocassidulina globosa* (Hantk.).

The latest Eocene species *Nodosaria coarctata* Hantk., *N. scalaris* Batsch., *Gyroidina suborbicularis* Katch., *Svratkina perlata* (Andreae), *Cibicidoides speciosus* (Cushm. et Ceder.), *Cassidulina oblonga* Reuss, *Cassidulinoides* sp., *Bulimina sculptilis* Cushm., *Bitubulogerina kasselensis* (Batjes), *Uvigerina mexicana* Nutt., *Angulogerina gracilis tenuistriata* (Reuss), *Baggatella* sp., and *Loxostomoides teretum* Cushm. are characteristic as well of the Oligocene deposits.

In terrigenous Oligocene deposits, benthic assemblages become impoverished and different in species composition. Appearing in these facies are *Ammodiscus pelucidus* Andreae, *Spiroplectammina carinata* Orb., *Spiroloculina canaliculata* Orb., *Plectofrondicularia vaugani* Cushm., *Valvulineria petroleyi* (Andreae), *Vsegeina* sp. nov., *Rotaliatina mamillata* (Andreae), *Herolandenia* sp., *Queraltina hispanica* Marie, *Pseudoplanulinella hierogliphica* Sigal, *Epistomina* aff. *paleogenica* Mjatl., *Asterigerina carinata* Orb., *Korobkovella* sp., *Pararotalia zeivensis* (Chal.), *Cassidulina laevigata* Orb., *Globobulimina socialis* Born., *Uvigerina taberana* Berm., *Loxostomoides sinuosum* Cushm., *L. vicksburgensis* (Cushm.), *Bolivina antegressa minor* Chal., *B. melettica* Andreae, *Chilostomelloides normalis* Chal., and *Sphaerogypsina* cf. *globulus* (Reuss). The most significant change in the benthic assemblages takes place at the base of the *G. tapuriensis* Zone. Accordingly, the Eocene–Oligocene boundary recognized on the basis of planktonic and benthic assemblages is at the same level.

In the northern areas, where no diverse planktonic assemblages were found, the position of the upper Eocene boundary may vary. As stated above, the *Nonion curviseptum* and *Almaena taurica* beds are referred to the Eocene. In the northern Caucasus (the Kheu River section), the *N. curviseptum* Beds contain the typical Eocene species *Hantkenina suprasuturalis* Bronn. (table). In the Kuban' River section, the topmost calcareous clay (4 m thick) of the Belaya Glina Horizon yields *N. curviseptum* in association with the late Eocene species *Globigerina gortanii* Borss. and *Turborotalia centralis* (Cushm. et Berm.). At the same level, Tabachnikova recognized the *Discoaster barbadiensis* Zone.

In the Crimea (the Kyzyl-Dzhar section), 25.75% of 66 established species are confined to the *A. taurica* Beds, and 65.15% appear in the underlying deposits being unknown in higher levels. A minor part of the assemblage consists of transient and late Eocene species, the latter passing into the Oligocene. Analyzing her own materials from the same section, Pechenkina (1971b) estimated the comparable proportions between species (23 and 58.9% respectively).

In the southern Emba area (Borehole 448-Tengiz), the upper part of the section (interval 45–51 m) is barren of planktonic foraminifers. Benthic forms are represented here by *Planulina costata, Bolivina antegressa*, and *Nonion curviseptum*. According to Tabachnikova, the mentioned interval yields nannoplankton of the NP21 Zone. Spherical tests of radiolarians, spines of echinoids, fish scales and rounded otoliths were also found. Abundant juvenile forms, tests of abnormal structure, and worm burrows evidence for unfavorable conditions for benthic foraminifers. The Eocene–Oligocene boundary is drawn at the depth level of 35 m within the greenish gray clays, where the Oligocene planktonic forms *Globigerina khadumica* N. Byk. and *Tenuitella liverowskayae* (N. Byk.) appear. This also the first occurrence level of *Glomospirella kugultinskensis* Schutzk., *Textularia agglutinans* Orb., *Cibicidoides kugultaensis* (Schutzk.), and *C. extremus* (Schutzk.) in the benthic assemblage. Nannoplankton was not found here. The depth level of 30 m yielded the recurrent fauna, ten species of the *P. costata* Zone, *Bolivina antegressa* included, and also the reworked globotruncanid tests and other Late Cretaceous forms.

In the northeastern part of the basin (northern Turkmenistan, the southern Aral area), where similar facies are devoid of plankton, the Eocene–Oligocene boundary is drawn at the level of changes in benthic assemblages. The assemblage of the *P. costata* Zone from the upper half of the Akhchakaya Formation is replaced upward by the assemblage of "*Lenticulina*" aff. *herrmanni* (with nonlaciniate peripheral margin), typical "*L.*" *herrmanni* (Andreae), *Spiroloculina canaliculata* Orb., *Bolivina mississippiensis* Cushm., *Loithostomata macrura* (N. Byk.), and other species characteristic of the Oligocene Daudanskii Formation. *Brotzenella munda* (N. Byk.) occasionally appears among these species (Borehole 4, Kunya-Urgench).

In the northwestern Ustyurt, the upper Eocene section is crowned by white marls bearing *Bolivina antegressa* and *Nonion curviseptum* (Nikolaeva and Prusova, 1979). The overlying clays with fish and *Planorbella* remains contain smaller planktonic foraminifers. To the northeast, deposits of the Chegan facies affinity yield benthic foraminifers known from the upper Eocene and passing into the Oligocene. These are *Melonis dosularensis* (Chal.), *Svratkina perlata* (Andreae), *Heterolepa almaensis* (Schutzk.), and *Cibicidoides sulzensis* (Herrmann) associated with rare *Brotzenella munda* (N. Byk.) appearing at this level. The cited authors attribute this fauna to the Oligocene, and I should agree with them.

Comprehensive investigations are needed to elucidate the boundary position in the northern Aral and northern Ustyurt sections characterizing the basin

Distribution of foraminifers near the Eocene–Oligocene boundary in the Kheu River section

Series	Eocene					Oligocene			
Horizon	Belaya Glina					Pshekha			
Thickness, m	0.5	0.5	0.2	0.3		0.1	0.3	0.3	
Species \ Sample number	1	2	3	4	5	6	7	8	9
Hantkenina suprasuturalis									
Globigerina officinalis									
G. ouachitaensis									
G. galavisi									
G. ampliapertura									
Globanomalina aff. *micra*									
Globorotalia postcretacea									
Chiloguembelina cubensis									
Planulina costata									
Bolivina antegressa									
Nonion curviseptum									
Turkmeniella kubynii									
Uvigerina eocena									
Bulimina truncana									
Svratkina perlata									
Vsegeina nana									
Tergrigorjanzeella sectile									
Ceratobulimina intrusa									
Heterolepa pygmea									
Eponides umbonatus									
Angulogerina angulosa									
Haplophragmoides stavropolensis									
Gaudryinopsis gracilis									

Note: Solid lines show the species distribution ranges in the section. Sample 4 is collected from the pyritized interbed; Sample 7 yielded the reworked tests of foraminifers *Acarinina bullbrooki*, *Globorotalia caucasica*, and *Globotruncana* sp. typical of the Belaya Glina Horizon.

facies that accumulated under cooler environments. Peculiar foraminifer assemblages from these sections blur the general trend of faunal changes throughout the section. T.P. Bondareva placed the Oligocene lower boundary above the *Bolivina antegressa* Zone at the base of the "*Anomalina munda*" (=*Brotzenella munda*) Zone. The typical marine fauna of the former zone includes representatives of relatively deep-water genera (*Bulimina, Uvigerina, Bolivina*, and others), whereas the *B. munda* Zone is characterized by the shallow-water Nonionida and Miliolida forms adapted to marine settings with unstable salinity.

Later, the *B. munda* s.l. Zone was subdivided into two parts (*Maikopskie...*, 1964). The lower one recognized as "*Cibicides salensis*" Beds yields species characteristic of upper beds of the *Bolivina antegressa* Zone. These are *Melonis dosularensis* (Chal.), *Fursenkoina hantkeni* (Cushm.), *Ceratobulimina intrusa* N. Byk. =? *C. Contraria* (Reuss), *Tergrigorjanzaella sectile* (Ter-Grig.), Miliolida forms, and *Brotzenella munda*. The upper part of the *B. munda* s.l. Zone corresponds to the "*Cibicides macrurus (=Loithostomata macrura)*" Beds bearing abundant tests of the zonal index species (acme zone) in association with *Loithostomata macrura* (N. Byk.) unknown from Eocene deposits and with typical Oligocene species *Asianella subbotinae* (N. Byk.) and *Uvigerinella* ex gr. *californica* Cushm. The Eocene–Oligocene boundary could be placed in this case at the base of the *L. macrura* beds, i.e., inside the *B. munda* s.l. Zone.

Other researchers who studied the northern Ustyurt and Chagrai Plateau recorded similar changes in com-

position of foraminifer assemblages. So, the foraminiferal assemblage from the lower part of the Chegan Formation is actually typical of the *Bolivina antegressa* Subzone (data of Bryzzheva in Bronevoi *et al.*, 1967). Some of its species pass into the overlying deposits, where *Tergrigorjanzaella sectile* (Ter-Grig.), *Brotzenella munda* (N. Byk.), and Miliolida forms characteristic of upper beds in the section, e.g., *Triloculina enoplostoma* (Reuss) and four species of the genus *Quinqueloculina*, were found. "There is a very gradual transition from the lower assemblage of the Chegan Formation to the upper one and, therefore, the boundary between them is conventional" (Bronevoi *et al.*, 1967, p. 39). The assemblage from the upper part of the formation was attributed by these researchers to the *B. munda* Zone, but I think that it may correspond only to the "*C. salensis* Beds" of this zone.

A similar assemblage enriched in arenaceous foraminifers was recovered from the lower part of the Turangli Formation of the northern Aral area (Bronevoi *et al.*, 1967, p. 56). The upper part of the formation corresponds to the first occurrence level of *Loithostomata macrura*, and the basal part of the overlying Ashcheairyk Formation yielded the assemblage of "*Lenticulina" herrmanni* (Andreae), *Bolivina mississippiensis* Cushm., and *Loithostomata macrura* (N. Byk.), which are definitely of the Oligocene age.

The analysis of published data shows that foraminifers from the considered stratigraphic interval should be studied further.

REGIONAL FORAMINIFERAL ASSEMBLAGES

Distribution areas of the studied late Eocene foraminifers represent parts of a single Belaya Glina basin that extended from the Carpathians to the Aral–Turgai area and eastern Central Asia (the Tajik and Fergana bays). Everywhere in the basin, biota evolved under subtropical conditions, as palynological data show that subtropical flora grew even in the Turgai area. As a consequence, differences between foraminiferal assemblages of the Belaya Glina reflect influence of the facies rather than climatic factors. The basin was connected with shallow inner seas of the Transcaucasia, assemblages of which reflected their proximity to the tropical Mediterranean region.

The Crimea. The assemblage of benthic foraminifers from the Al'ma Horizon is most representative in the Crimea–Caucasus region. It includes 77 genera and more than 140 species (Bugrova, 1988a), most of which are widespread outside the Crimea. Calcareous forms are dominant (64 genera and 120 species), whereas the abundance rate of arenaceous foraminifers that slightly increases upward in the section is up to 10–12%. Many genera are represented by one or two species in contrast to other more diverse genera *Robulus* (not less than 10 species), *Marginulina* (7), *Cibicidoides* (6), *Planulina* (3), *Bulimina* (4), *Uvigerina* (6), and *Bolivina* (4). According to its composition, this fauna evolved in the epicontinental sea that was connected with the Tethyan and North European (Polish) basins.

The northern Caucasus and Ciscaucasia. Assemblages from moderately deep-water deposits in the Kuban' and Kheu river basins are identical in taxonomic composition to those from the Crimea. The Kheu River section is characterized by 149 species of 79 genera. Calcareous foraminifers are still dominant here, but abundance of arenaceous forms is higher (22 species, or 14.7%). The impoverished assemblage from more clayey facies of the Stavropol' area shows a higher proportion of arenaceous forms. In total, 52 species of 29 genera are known from this area (Ter-Grigoryants, 1965), and 14 of them are agglutinating. In the Ipatovo area located northward, only 34 species of 24 genera were found (collection of V.T. Balakhmatova from the reference borehole).

The Russian platform and the Scythian plate. Fauna from this shallow-water part of the basin is less diverse than that of the Crimea and Caucasus. In the Volga area near Volgograd (Subbotina, 1953), 61 species of 39 genera were found. Slightly calcareous clays of the Volga–Don interfluve (the Melikhov hydrosystem) yielded 71 species of 51 genera, and 12.5% of these species have tests with arenaceous walls; their share increases upward in the section. The fauna mostly consists of the Crimean–Caucasian species accompanied by forms representing genera *Plectofrondicularia, Spirobulimina, Neoeponidella, Astrononion*, and others. Most abundant tests are those of *Robulus, Bulimina*, and *Uvigerina* genera. Agglutinating foraminifers were encountered only in coastal shallow-water deposits of the Voronezh anteclise (*Geologicheskie...*, 1996). *Cyclammina pseudocancellata* (the index-species of the local zone) is abundant in lower beds of exposures in the Don River lower courses and in the Karpinskii Ridge (Nikitina, 1963, 1972). The local biostratigraphic unit bearing *Neoeponidella lucida* (*Asterigerina* in terms of Nikitina) and *Cibicidoides salensis* was recognized in the Sal–Manych interfluve. Beds with these species were included into the *Bolivina antegressa* Zone of the Azov–Volga–Don region (*Geologicheskie...*, 1996). According to E.Ya. Kraeva and I.D. Konenkova, a similar assemblage occurs in the Black Sea depression. Nearly 60 species of 35 genera, those of shallow-water genera *Asterigerina* and *Rotalia* included, were found in the Odessa–Kherson area. Some species of the *P. costata* Zone are widespread in Moldavia (*Geologicheskie...*, 1996; collection of V.T. Balakhmatova).

The Transcaspian areas of Central Asia. In the Krasnovodsk Peninsula of western Turkmenistan, clayey marls and carbonate clays contain 88 species of 45 genera, dominant among which are anomalinid; representatives of eight genera have arenaceous tests. The assemblage of the *P. costata* Zone is recorded in the

eastern coast of the Kara-Bogaz-Gol Bay, in central and northern Turkmenistan (Tyzkyr, 43 species of 34 genera), in the Zaunguz area of the Kara-Kum, and in the southern coast of the Aral Sea (82 species and 51 genera). In the last area (Borehole 1-Takhta), the pyritized interbed at the top of the Eocene sequence bears yielded larger *Bulimina* and *Uvigerina* tests, and *Tergrigorjanzaella sectile* appears in the overlying Oligocene beds (*Reshenie...*, 1970).

More than 30 species of the Belaya Glina Horizon unassociated with endemic forms were found in marls and carbonate clays of the eastern Ustyurt Plateau, but only some of them were encountered in individual interbeds of the noncarbonate Kyzyl-Kum sections (*Geologicheskie...*, 1996; and publications of N.V. Averburg).

Sections of the West Turkmenistan mountains are composed of siliceous clays containing impoverished foraminiferal assemblages. The middle Torymbeur Subformation of the Malyi Balkhan yielded 35 species of 26 genera (collection of L.S. Nevzorova). This assemblage includes 26 species of 22 genera known in the Belaya Glina fauna, but associated species *Textularia hockleyensis* and genera *Rotalia, Astrononion, Florilus*, and *Baggatella* are unknown in that fauna. The lower beds of the section contain some elements typical of Central Asia, whereas the upper ones show an increasing number of species characteristic of the Belaya Glina Horizon, most abundant of which are *Bulimina* and *Uvigerina* forms. A similar assemblage was discovered in the western Kopetdag (collection of M.G. Zakharova), where the middle Torymbeur Subformation yielded *Tergrigorjanzaella sectile* (Ter-Grig.) (N.K. Bykova, personal communication). To the east, shallow-water deposits of the Gyaurdag contain an impoverished assemblage of about 20 species, the endemic forms and species of the Tajik—Fergana region included (data of G.E. Kozhevnikova).

Upper Eocene deposits occur in localized areas of southeastern Turkmenistan. Assemblages are less diverse here despite the fact that sections are composed of calcareous clays. An assemblage from middle courses of the Amu Darya River consists of 50 species of 41 genera characteristic of the Belaya Glina Horizon (collection of Balakhmatova). In the Mary area, 20 species of this kind associate with *Brotzenella vakhshica* characteristic of eastern areas, but endemics are unknown here.

The foraminiferal assemblage of the Tajik region is representative. According to Davidzon *et al.* (1982), it includes as many as 73 species and forms belonging to 31 genera. About 41% of the species are identical to those of the Crimea–Caucasus region, and 17% of the species are endemics, predominantly the shallow-water arenaceous forms (genera *Trochammina, Textularia*, and *Haplophragmoides*). Assemblages from the region demonstrate compositional changes depending on the host-rock facies. In the west, assemblages from marls deposited in an open sea-bay setting are most similar in composition to the fauna of the *P. costata* Zone. For example, only two of 40 species (representing 24 genera) from the Gaurdak-Kugitang area are endemic. A similar assemblage is known from the Gissar–Karateg area, but in the easterly Kafirnigan area, terrigenous sediments contain the impoverished fauna, which includes some primitive arenaceous forms (Davidzon *et al.*, 1982).

Foraminiferal assemblages from coastal deposits of the past Tajik and Fergana bays are of a simple taxonomic composition. In the Khanabad Horizon of the Fergana Bay, only several species, e.g., *Brotzenella vakhshica* (Davidz.) and *Cibicidoides karatauensis* Davidz. are in common with those from Tajikistan (*Reshenie...*, 1970). The upper part of the indicated horizon yields *Neoeponidella lucida* (Minak.) in association with *Globigerina postcretacea* Mjatl. The overlying beds (Sumsar Horizon) enclose the Oligocene planktonic species *Globigerina khadumica* N. Byk., *Tenuitella liverowskayae* (N. Byk.) and benthic forms *Brotzenella munda* (N. Byk.) and *Loithostomata macrura* (N. Byk.), the latter widespread in the Aral Sea area and Ustyurt.

Kazakhstan. Foraminiferal assemblages from different parts of the respective ancient basin vary in composition and structure. In carbonate deposits of the western Ustyurt chinks, Mangyshlak, and southern Emba area, the assemblages includes species of the *P. costata* Zone and a series of endemic forms. In her manuscript, Bykova defined the latter adding the remark "var." and her identifications should be certainly revised. The Mangyshlak deposits yielded 50 species of 37 genera, most abundant of which are *Bulimina* and *Uvigerina* forms. In the western Ustyurt chinks, 44 species of 27 genera were found in the Bayurbas sections (Bykova, 1974), 57 species of 37 genera in the Tuzbair sections (Subbotina, 1953), and 57 species of 36 genera in the Kenderly sections. *Nonion curviseptum* was detected here in the uppermost Eocene beds.

Calcareous clays of the southern Emba region (Borehole 448-Tengiz) bear 52 species of 34 genera. According to unpublished data of Bykova, this regional assemblage is actually more diverse and consists of 79 forms (including those with remark "var.") representing 46 genera.

In the northwestern Ustyurt (Nikolaeva and Prusova, 1979), marls and calcareous clays are characterized by the Belaya Glina assemblage, and the uppermost Eocene beds yield abundant specimens of *Nonion curviseptum* and *Bolivina antegressa*. The overlying Oligocene clays contain fish remains and *Planorbella* forms. In the northern Ustyurt chinks (Bronevoi *et al.*, 1967; collections of Bryzzheva), the assemblage of 36 species and 23 genera from the lower Chegan Subformation is similar to the Crimean–Caucasian assemblages, and 76.6% of 30 species from Borehole 509 are

common for two regions as is mentioned in the publication. The uppermost Eocene beds yielded an impoverished assemblage (24 species of 20 genera) that includes nonionid forms and *Tergrigorjanzaella sectile* known from the northern Caucasus, Russian platform, and Poland. Nonionida and Miliolida represent the shallow-water groups adapted to basinal settings with lowered temperature and salinity.

The northernmost distribution area of the Belaya Glina species is the Turgai depression. The Turgai assemblage consists of 11 species of 11 genera. Among them (collections of I.M. Aizenshtat and N.I. Komarova), there are *Bolivinopsis tuaevi* Moroz., *Marginulinopsis infracompressa* (Thalm.), *Anomalinoides affinis* (Hantk.), *Baggina iphigenia* (Samoil.), *Bulimina truncana* Guemb., *Bolivinoides reticulatus* Hantk., *Epistomina paleogenica* Mjatl., and abundant *Cyclammina pseudocancellata* Chal. Aizenshtat distinguished the local *Cyclammina pseudocancellata* Zone in the lower Chegan Subformation. Data on endemics have not been reported, but the fauna in question could be inadequately studied. It is clear, nevertheless, that the Turan basin was connected with the Caucasian basin and, to a lesser extent, with the Central Asian sea. Fauna of the basin developed under influence of biota that inhabited the southwestern sea areas.

The Transcaucasia. Georgia. Regional assemblages of benthic foraminifers are close in composition to the Belaya Glina fauna being simultaneously less diverse. As is reported by Kacharava (1977), benthic foraminifers occur in marls and marlaceous clays of the northern slope of the Trialeti Ridge (14 species, 12 genera) and in the Akhaltsikhi depression (24 species, 21 genera). These and the Karatuban' sections, the latter studied by myself, contain all characteristic species of the *P. costata* Zone, including *Nonion curviseptum* found in the uppermost Eocene beds. The assemblage is dominated by secreting forms (genera *Robulus, Heterolepa*, and *Bulimina*). Endemics are scarce, because the mentioned species are widespread outside the region. Probably, this basin was connected more closely with northern sea basins than with southern ones.

Azerbaijan. Foraminiferal assemblages from the eastern Caucasus and Transcaucasia are similar in general to the Belaya Glina fauna (Khalilov, 1962). The similarity is better pronounced in northeastern sections than in southwestern ones containing some Mediterranean elements (genera *Planorbulina, Halkyardia*, and *Rotalia*). Abundant *Nonion, Bolivina*, Anomalinidae, and *Chilostomella* species, predominantly their endemic varieties, occur everywhere. The uppermost Eocene beds of Azerbaijan yield *Nonion curviseptum.*

In marls and marlaceous clays of northeastern Azerbaijan, predominantly calcareous tests of 29 species (17 genera) were found. Assemblages from the Caspian area are more diverse (40 species of 20 genera) and include the genus *Pyrgo* widespread in the Mediterranean region. Fauna from piedmont areas of the Lesser Caucasus is peculiar in composition (83 species of 41 genera). It includes scarce agglutinating forms (6 mostly primitive genera) and diverse representatives of genera *Bolivina* (17 species), *Nonion* (6 species), *Chilostomelloides* (5 species) *Chilostomella* (3 species), and *Rotalia* (2 species). Many species of these and some other genera (especially anomalinids) are endemic. The assemblage from the Kobystan locality (84 species of 41 genera) includes the Tethyan genera *Articulina, Halkyardia, Planorbulina*, and *Rotalia* (4 species) unknown from other areas. All data suggest that the shallow-water inner basins of the Transcaucasia were connected with the Mediterranean basin.

Armenia (sections of the Landzhar and western Aiotsdzor sites; materials of L.A. Pankova and K.N. Paffengolts). Deposits of the mentioned sites yielded 134 species of 76 genera. About 25 species represent the warm-water genera *Pseudoplanulina, Queraltina, Halkyardia, Schlosserina, Chapmanina, Pyrgo*, and some others unknown from the northern areas but typical of the South European sections. The similar assemblages of nummulitids, benthic and planktonic foraminifers clearly indicate the Mediterranean-type affinity of respective biota that evolved in a single basin.

PALEOGEOGRAPHIC POSITION OF THE BELAYA GLINA BASIN

The composition of benthic foraminifers from the Belaya Glina basin suggests that the basin was a part of the Mediterranean paleobiogeographic realm. Their assemblages are lacking characteristic elements of the North European fauna. The latter is exemplified by the Lattorfian assemblage from northern Germany (Kiesel and Lotsch, 1963), where it includes other characteristic genera: *Lacazina, Tubulogerina* (3 species), *Reussella* (5 species), *Conorbina, Biapertorbis, Articulina, Sphaerogypsina, Bolivinella* (2 species), *Planorbulina, Ceratobulimina*, and *Asterigerina.* The Belaya Glina basin was connected with the North European basin only via a shallow strait formerly located in the plain area of Poland, where fauna with *Tergrigorjanzaella sectile* and other characteristic species of the *P. costata* Zone is recorded (Pozaryska, 1978).

The Belaya Glina assemblages contain species and genera known in the Mediterranean region. For example, among 57 genera and nearly 90 species reported from the Priabonian stratotype area (Northern Italy), most of the former and not less than 20 species have their counterparts in the Belaya Glina fauna. (It is appropriate to mention here possible differences in species identifications, as one can judge from published pictures of specimens).

Coastal exposures in the Maritime Alps yielded 58 species of 43 genera (Bodelle *et al.*, 1968; and others). Despite the more thermophilic character of this assem-

blage in general, 24 its species are identical to those of the Belaya Glina fauna. All genera and not less than 16 species in assemblages from the Navarro and Catalonia provinces of Spain (altogether 55 species of 38 genera; Ruiz de Gaona and Colom, 1950) have analogs in the Belaya Glina assemblage. Many forms characteristic of the Belaya Glina fauna were described by Hantken from Hungary and by Guembel from the Alpine sections of Bavaria. Finally, the Armenian fauna similar in composition to the Mediterranean one also contains some characteristic Belaya Glina species.

The distinct similarity between the indicated faunas (despite the all shortcomings in investigation and identification of taxa) suggests that the Belaya Glina basin represented the northern province of the Mediterranean paleobiogeographic realm. The province extended over a part of the future Paratethys from the Carpathians and the Black Sea region to Central Asia.

The Belaya Glina plankton and benthos colonized a series of depressions in the eastern periphery of the province (the southeast of Turkmenistan, Tajikistan, and Fergana area). Assemblages with *Brotzenella vakhshica* and abundant endemics were characteristic of these depressions. Endemic forms of genera *Nonion* and *Nonionella* populated shallow settings with low water salinity together with anomalinids and arenaceous foraminifers representing genera *Haplophragmoides, Textularia, Bolivinopsis*, and *Trochammina*. Fauna of this type is used to distinguish the Tajik subprovince approximately corresponding to the distribution area of *Brotzenella vakhshica*. Currently available data on foraminifers from the Aral–Turgai region are insufficient for considering it as an independent paleobiogeographic subdivision.

Fauna of the inner seas of Azerbaijan includes many Belaya Glina species but demonstrates a high degree of endemism. The genera *Chilostomella* and *Chilostomelloides* and some species of genera *Bolivina, Anomalinoides, Heterolepa*, and *Nonion* are unknown outside these seas. The transitional assemblage from the Kobystan predominantly consisting of the Belaya Glina taxa also includes genera *Articulina, Planorbulina*, and *Halkyardia*. Evidently, the respective basin was connected with the Mediterranean Sea via the Armenian bay and with the Belaya Glina basin via the Caspian area in northeast. According to these biota peculiarities, the eastern Subprovince of Transcaucasia can be recognized.

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Transgressive and Regressive Seismic Facies of the Danube Deep-sea Paleofan

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Abstract—Based on the seismostratigraphic analysis and geological interpretation of seismic profiling data, three marker surfaces corresponding to structural unconformities and erosion events are defined in the area of the Danube and Dnieper deep-sea fans in the Black Sea continental slope. It is established that sedimentary complexes concurrent to Quaternary glaciations correspond to three main phases in the Black Sea regressions and transgressions.

Key words: Black Sea, Pleistocene, deep-sea fan, regression, transgression, seismostratigraphy.

INTRODUCTION

Beginning from the 1960s, extensive seismostratigraphic investigations with the help of common midpoint (CMP) and continuous seismic profiling methods were carried out in the distribution area of the giant Danube deep-sea paleofan in the western and northwestern parts of the Black Sea basin. In these investigations, air guns and sparkers were used as the seismic wave sources. Seismic researches with application of the CMP method were mainly aimed at the study of the entire thick sedimentary sequence. In this connection, the low-frequency seismic profiling with a relatively low resolution was used. In the seismic profiling sections across the paleofan, the sedimentary body proper represents only an insignificant part of the latter thus providing opportunity to infer only the general information on the form, dimensions, and spatial position of the deep-sea fan on the continental slope of the Black Sea.

First seismoacoustic study of the Danube deep-sea paleofan was carried out in 1969 from the R/V "Atlantis-II" (Ross *et al.*, 1974). The study provided first data on the paleodelta dimensions. Main parameters and general structural features of the latter were established during the CMP survey during the subsequent regional seismic investigations performed by the Research–Industrial Association "Yuzhmorgeologiya" in 1975–1976 (Kazantsev *et al.*, 1978; Tugolesov *et al.*, 1985; Gorshkov *et al.*, 1993).

The detailed data on the inner structure of deep-sea paleofans of the Danube, Dniester, and Dnieper rivers spread over the northwestern slope of the Black Sea were obtained during the international expedition on the R/V "Gelendzhik" organized by the State Organization of the Central Geological–Geophysical Expedition ("Goskomnedra") in 1993 (Shimkus *et al.*, 1997). Scientists from Romania and Germany took part in this cruise along with Russian researchers. The data obtained enabled the seismostratigraphic analysis of the intricate structure of the sedimentary sequence in the considered deep-sea fan. These materials served as the basis for the present paper.

The works were carried out, using the single-channel seismic profiling reflection method in the area 110 × 100 km in size along 12 parallel profiles with interjacent distance of 10–11 km and along 4 transverse profiles (Fig. 1). The profiles cross the accumulative body of the deep-sea fan across its strike. The study area stretches from the shelf edge to the foot of the continental slope and further to the abyssal part of the Black Sea basin. The area comprises main parts of the fan: incisions of paleovalleys near the shelf edge and accumulative bodies of the median and distal parts of the deep-sea fan.

The high-frequency seismic studies provided the first data on the inner structure of the Danube deep-sea fan.

In addition to these new materials, the results of seismic CMP investigations obtained by the Research–Industrial Association "Yuzhmorgeologiya" (1975–1976), Southern Division of the Institute of Oceanology, Russian Academy of Sciences (R/V "Professor Shtokman," 1981) (Kogan *et al.*, 1987), and by the Geophysical Institute of Trieste (Finnetti *et al.*, 1988) were used.

The deep-sea fan of the Danube, Dniester, and Dnieper rivers occupies almost the entire northwestern continental slope of the Black Sea and the southern part of its shelf with the distal part extending far into the abyssal part of the basin. It covers the area about

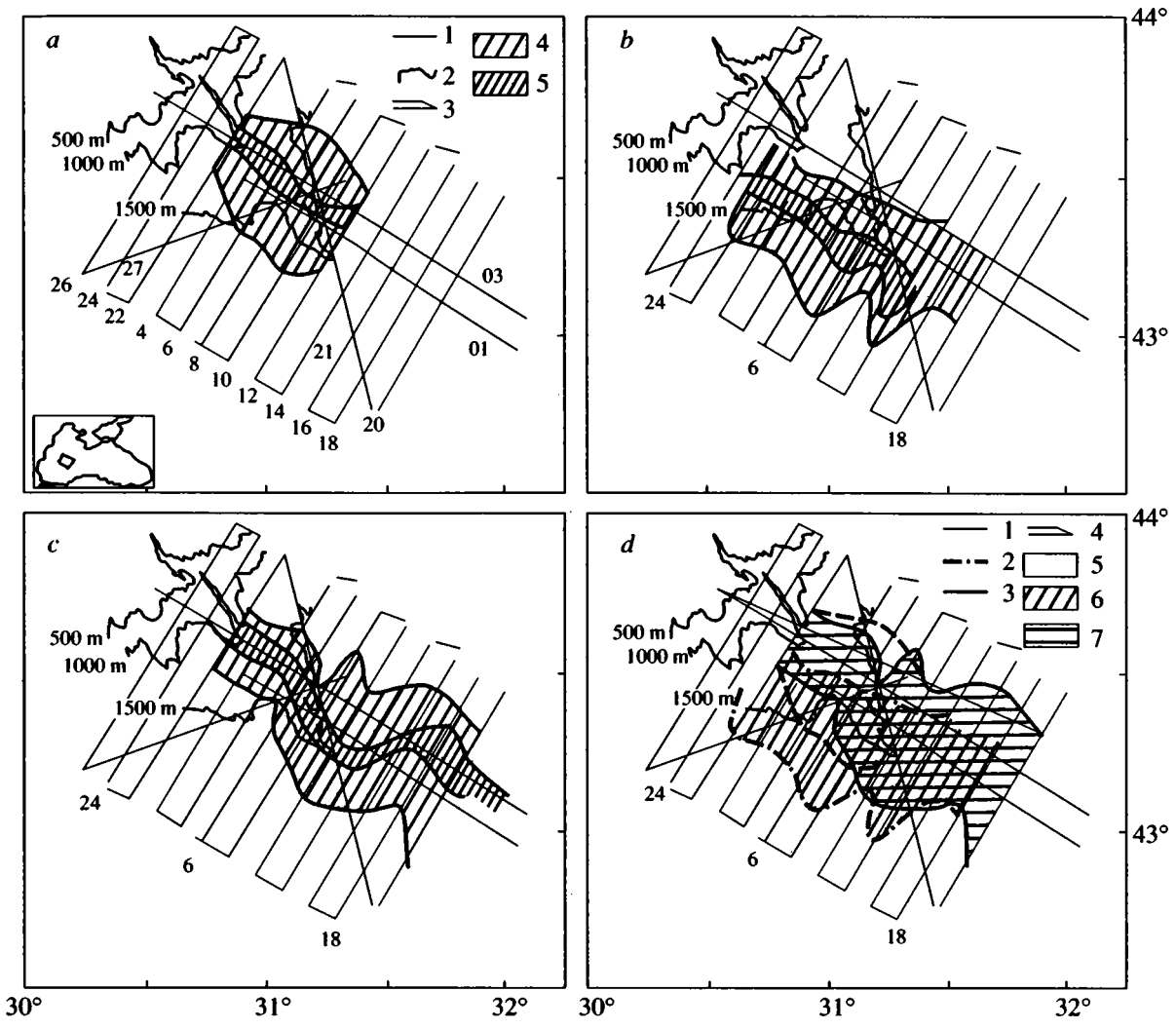

Fig. 1. Distribution areas of deep-sea fans formed during the Mindel (a), late Riss (b), and late Würm (c) epochs and their relative positions (d) in the study region; (*a*, *b*, *c*): (1) contours of sedimentary bodies, (2) depth contours, m; (3) seismic profiles, (4) fan areas, (5) axial fan parts (the channel and near-channel banks); (*d*) contours of sedimentary bodies (1–3), seismic profiles (4), and areas occupied by fans (5–7).

50000 km^2 in size, and the thickness of its deposits is 2 to 3 km and more (Gorshkov *et al.*, 1993). The seismic survey of 1993 covered area that stretches along the main channel of the paleodelta. The earlier CMP seismic investigations revealed smaller paleodelta channels located westward of the study area (Shimkus *et al.*, 1997).

The huge accumulative deep-sea fan occupying the entire northwestern slope of the Black Sea basin from the Crimea to Bulgaria represents an intricate structure consisting of many fans variable in size. The Danube fans proper are those located in the axial zone of this giant accumulative body and westward of the latter. The deep-sea paleofan formed by the Dnieper, Dniester and, probably, by some Crimean rivers is located eastward.

SEISMIC UNITS OF DEEP-SEA FANS

The transverse section of the deep-sea fan reveals easily defined sedimentary bodies located at different depths and having well-expressed triangular outlines (Fig. 2b). Being traceable throughout the entire study area, they represent the swell-shaped accumulative bodies. Every of these bodies is usually overlain by a thin sedimentary layer. The swell-shaped accumulative bodies represent buried deep-sea fans formed during one of the Black Sea regressions. The overlying sedimentary sequence accumulated under different conditions, probably during the subsequent sea transgression. Thus, the Danube deep-sea fan represents cyclic alternation of sedimentary sequences formed in different physicogeographical environments corresponding to sea transgressions and regressions. Sedimentary members accumulated during the sea transgression (the

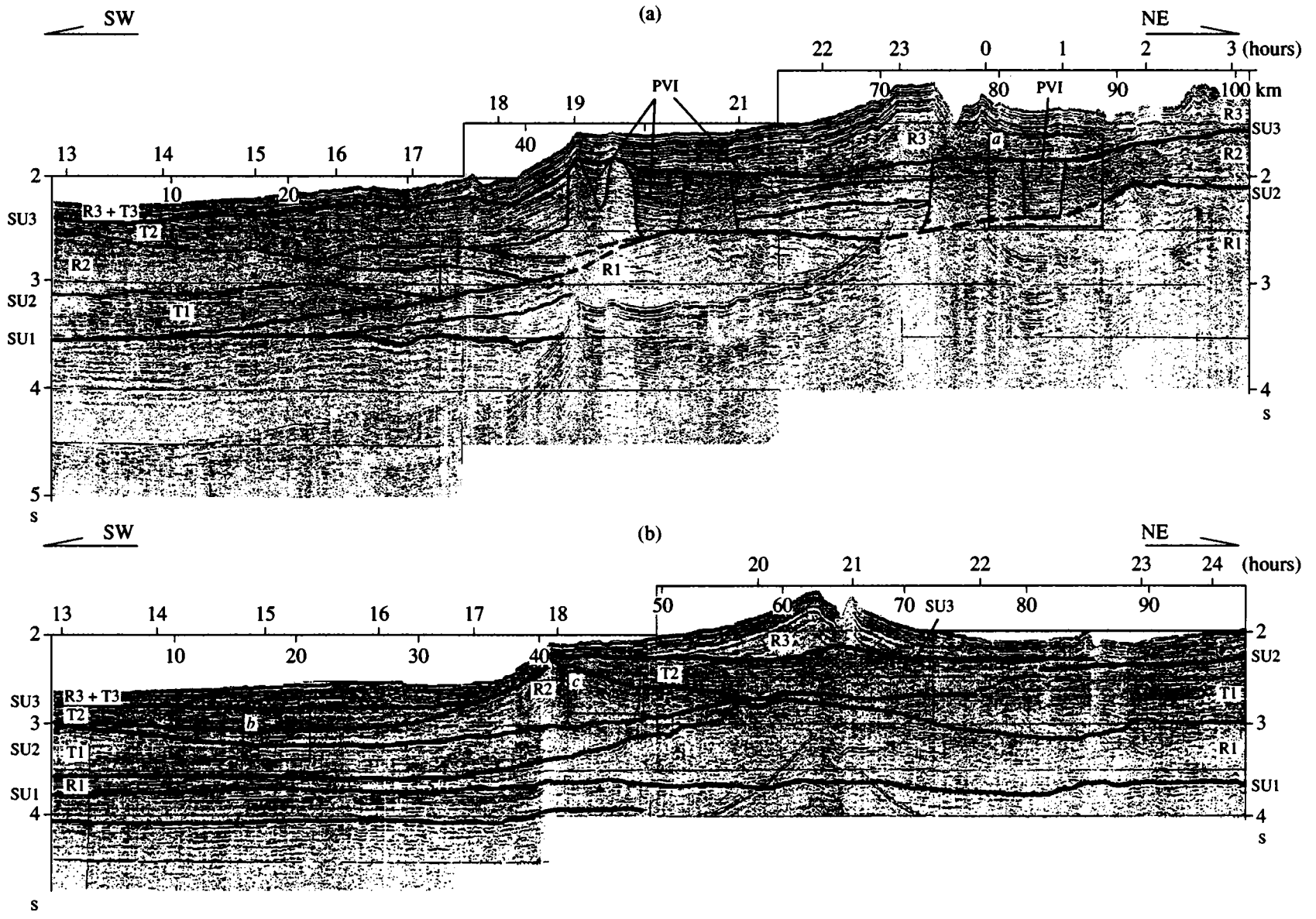

Fig. 2. Typical seismic records of the Danube deep-sea fan. (a) Profile 24; (b) profile 6; (c) profile 18. Position of profiles is shown in Fig. 1. Indices designate seismic units of the buried deep-sea fans: (SU1) lower, (SU2) middle, (SU3) upper; (T1–T3) transgressive sequences; (R1–R3) regressive sequences; (PVI) paleovalley incisions.

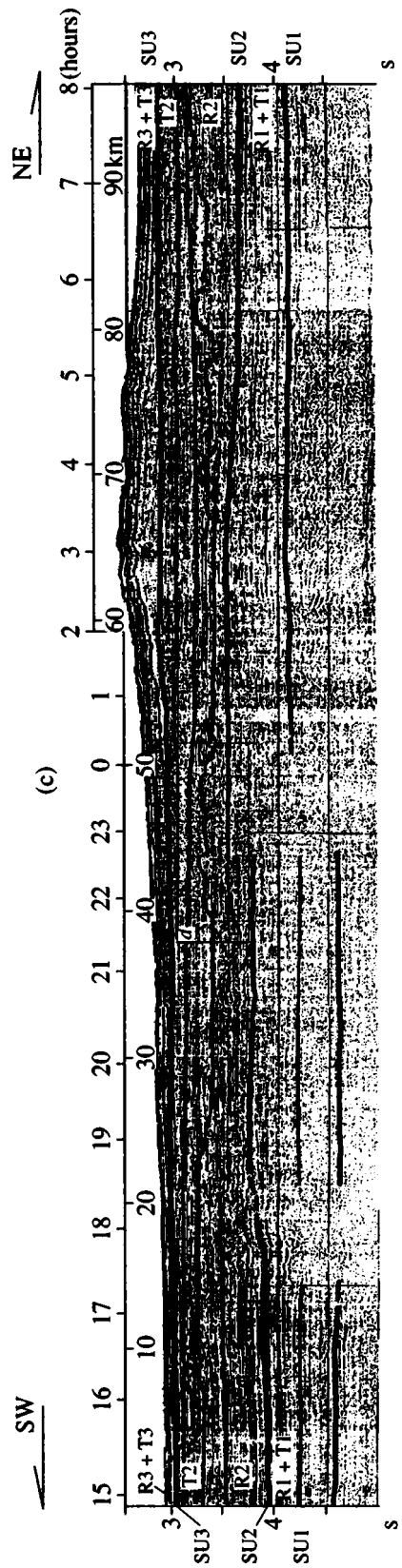

Fig. 2. (Contd.)

transgressive sequence) unconformably overly deep-sea paleofans formed during the regressive phase (the regressive sequence). The angular unconformity separates the transgressive sequence from the overlying sequence, which accumulated during the subsequent regression. The section of the Danube deep-sea fan in the study area comprises three regressive–transgressive cycles corresponding to three seismostratigraphic units. Every of these cycles includes a well-expressed swell-shaped accumulative body, which corresponds to the regressive phase of the sea history, and the overlying, usually bedded deposits of the transgressive phase (Fig. 2).

The sedimentary material is mainly concentrated in the lower unit. The upper unit is the thinnest one because it lacks the transgressive part, formation of which is still in progress.

Deep-sea fans located at the foot of the continental slope are gradually flattening out and grade into subhorizontal bedded stratiform sedimentary bodies (Fig. 2c). The younger accumulative bodies build up fans toward the deep basin. The sedimentary body of the fan corresponding to the upper unit is advanced by 45–50 km seaward relative to the fan of the lower unit.

In plan view, the summit of every fan resembles in morphology the meandering flow (Fig. 1). The surface of the upper unit, i.e., the modern bottom topography, was mapped using the side-view sonar (Shimkus *et al.*, 1977). It has outlines of a large river with meanders of a great amplitude. All three main deep-sea fans stretch along the axial part of the study area in the NW–SE direction. The summit of the middle unit is strongly shifted southward relative to other units. The shift amplitude is up to 20–25 km.

SEISMIC FACIES OF THE DEEP-SEA FAN

All three seismic units established in the sedimentary body of the Danube deep-sea fan reveal a uniform set of seismic facies recognized in both the regressive and transgressive sequences. Different seismic facies characterize sediments composing channel banks, the buried erosional valleys that are filled to a variable extent with sedimentary material, the subhorizontal sequences, discontinuously bedded and leaned against banks, and thin sediments with parallel bedding, which envelope both the channel-bank and leaned sequences.

Seismic facies of the regressive and transgressive sequences significantly differ from each other in the upper and middle parts of the fan. Only in the distal part of the latter, where the accumulative body of the regressive cycle is devoid of the channel-bank morphological features and shows a stratal shape, the seismic facies of the regressive and transgressive sequences become almost identical.

Most diverse are seismic facies of the upper seismic unit. This is probably a consequence of the applied seismic method, the resolution of which gradually decreases downward, smoothing geological outlines in the lower section part. The following peculiarities are characteristic of all three seismic units. Gradual facies transitions and replacements are established along the fan strike. In central parts of accumulative bodies representing a regressive cycle of every unit, seismic facies are usually characterized by most chaotic or small-scale hummocky patterns (Fig. 3c). Flanks of the ridges reveal sigmoid, clinoform, and bedded facies with the basal offlap (Fig. 3b). Transgressive sequences mainly represent subparallel, sigmoid, and hummocky seismic facies (Fig. 2b, c). The accumulative body of the regressive sequence in the upper part of the continental slope usually reveals several distinct reflectors traceable over a certain distance. In the middle part of the slope, the reflectors disappear, and the distal part of the accumulative body corresponds to thin-bedded and relatively ordered seismic facies (Fig. 3d).

Transgressive sequences of all three seismic units are virtually identical, characterizing sediments with subparallel bedding (Fig. 2c). Seismic facies similar in patterns grade into each other both along and across the fan strike.

Facies of the Upper Regressive Unit

The central part of the accumulative channel bank reveals the hummocky seismic facies (Fig. 3c). At the bank flanks, the hummocky facies grade into the facies with the subparallel bedding and basal offlap (Fig. 3b). In seismic records along profiles 10 and 12 (the central part of the study area approximately corresponding to the middle part of the channel bank), the inner seismic structure becomes less distinct, more chaotic, and more transparent. In the upper part of the channel bank, there are several seismic surfaces extending from the shelf edge downward the slope and dividing the sequence of the accumulative body into several beds (profiles 6 to 26). The upper bed (profiles 24 and 26) shows the bedded seismic facies, and the rest of the channel bank includes the hummocky and sigmoid facies. The bedded facies prevail in the distal part (Fig. 2c).

Facies of the Middle Regressive Unit

The small-scale hummocky and sigmoid facies are recorded in the central part of the accumulative body; they grade into the discontinuously bedded facies at the bank flanks. The facies replacement downward the slope is similar to that in the upper unit. The poorly expressed hummocky facies in the core of the upper (initial) part of the unit (profile 26) is replaced at the bank flanks by the small-scale and clinoform (sigmoid) facies. The central part of the accumulative body (profiles 12–22) corresponds to facies with chaotic patterns. Down the slope, beginning from profile 14, the latter grade into the hummocky facies. In the distal part of the accumulative channel bank, there are seen transitions from slightly stratified bedded facies to the small-scale hummocky or chaotic ones.

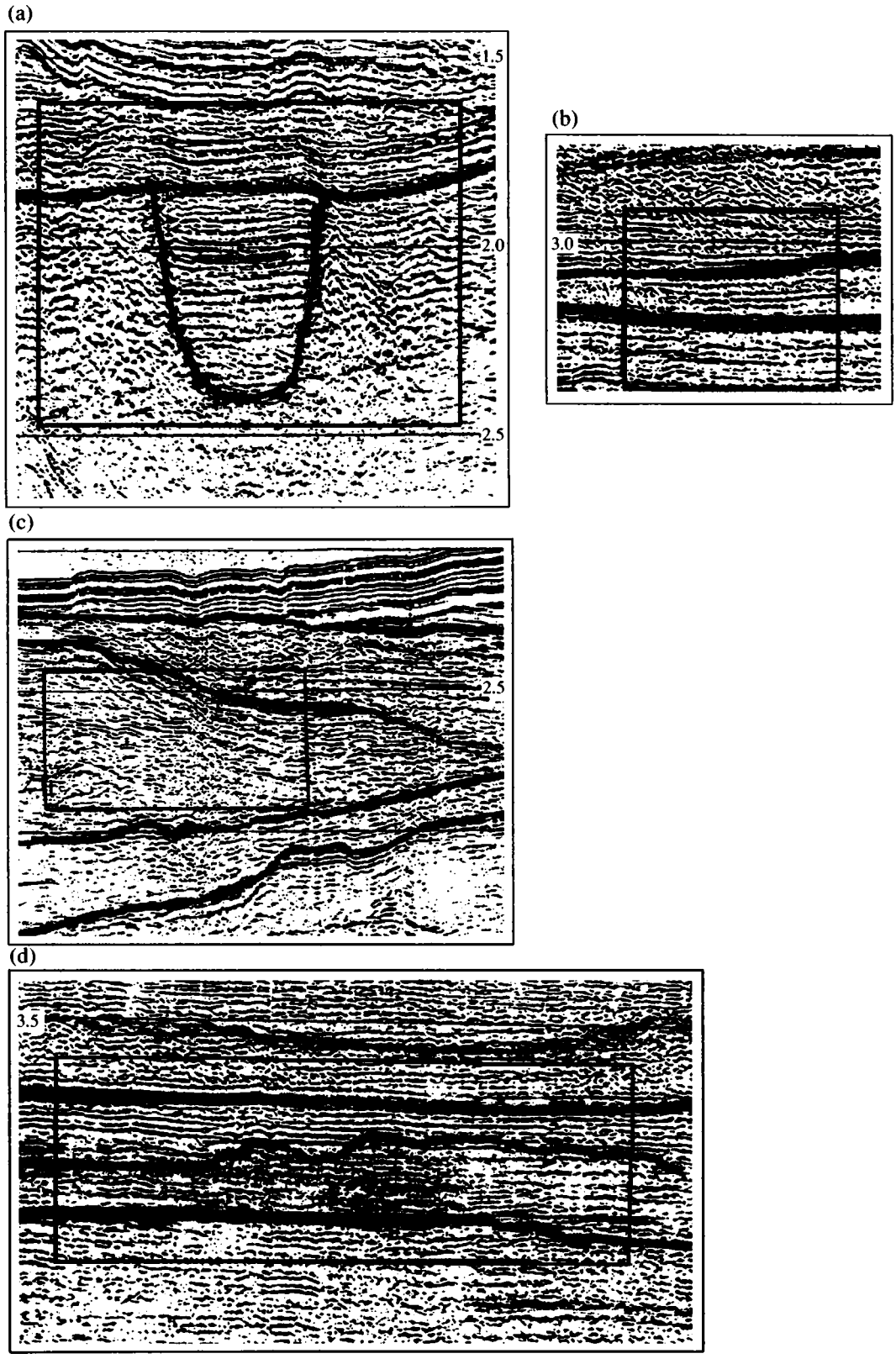

Fig. 3. Typical seismic facies of the deep-sea fan: (a) parallel bedding in sediments filling paleovalleys, (b) flank turbidite with parallel bedding in the middle part of the fan, (c) sigmoid–hummocky seismic patterns in the accumulative channel bank that includes a clinoform sequence in its middle part, (d) sigmoid–hummocky seismic patterns in the accumulative channel bank and discontinuous parallel (slightly chaotic) bedding in the distal part of the fan.

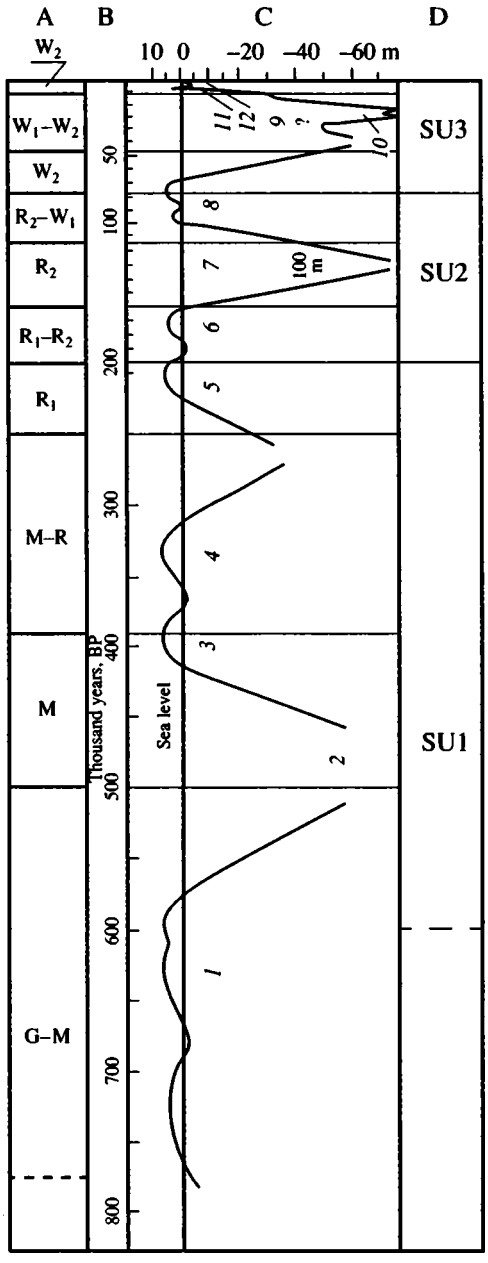

Fig. 4. Sea-level fluctuations in the Black Sea during the Quaternary, column (A): Würm (W), Riss (R), Mindel (M), and(G) Günz epochs; column (B) age, ka; column (C) secular sea-level fluctuations after Fedorov (1978): (*1*) Chaudan transgression, (*2*) post-Chaudan regression; (*3*) early Paleoeuxinian, (*4*) Paleouzunlarian, (*5*) late Paleoeuxinian, (*6*) Uzunlarian, (*7*) post-Uzunlarian, (*8*) Karangatian, and (*9*) Surozh transgressions; (*10*) post-Karangatian regression, (*11*) Novoeuxinian transgression, (*12*) Chernomorian (Holocene) transgression; column (D) ordinal numbers of deep-sea fans.

Facies of the Lower Regressive Unit

Dissimilar to overlying units, the core of the lower accumulative channel bank is characterized, along its almost entire length, mainly by chaotic and large-scale hummocky seismic facies, or by the obscurely bedded facies grading into the wavy bedded one at the bank flanks.

Facies of Erosional Submarine Valleys

Widespread in the uppermost part of the continental slope are erosional submarine valleys deeply incised into sediments of the upper and, partly, middle seismic units. Buried riverbeds also occur there inside the sequence of the Danube deep-sea fan. They are filled with sediments accumulated during the transgression peak. The sediments usually correspond to the seismic facies with thin parallel lamination (Fig. 3a).

TRANSGRESSIONS AND REGRESSIONS OF THE BLACK SEA

In the Black Sea, significant sea-level oscillations accompanied by significant regressions occurred since the Mindel glaciation (Fedorov, 1978, 1984). The major regressions, when the sea level dropped down by 60–100 m and more, occurred during the Mindel, late Riss, and late Würm epochs dated back to 500, 140, and 30 ka ago (Fig. 4) and corresponding to the Chaudan, Uzunlarian, and Karangatian glaciations, respectively. These three regressions in the Black Sea region are probably responsible for deposition of largest buried accumulative bodies detectable in seismic records across the Danube deep-sea fan and corresponding to three seismic units of its sedimentary sequence. The accumulation of the Danube deep-sea fan comprising thick accumulative bodies started approximately 1 Ma ago and continued through the entire glacial Pleistocene.

According to deep-sea drilling data (Ross *et al.*, 1978; *Geologicheskaya istoriya...*, 1980), the oldest glacial epoch (Günz) did not left any distinct signs of deep regression in the Black Sea. Regressions were distinct during the Mindel, Riss, and Würm glaciations in the period of the last 0.5 m.y. A similar inference was derived from the study of coastal sequences (Fedorov, 1978, 1984).

The accumulative bodies defined in seismic records correspond precisely to this time interval. During older periods, bodies of this kind either did not accumulate, or were localized northward of the study area, where they are overlain by the sequence of fluvial sediments of avandelta (within the modern shelf). This agrees with the fact that the boundary of the deep-sea basin was displaced significantly to the north from its modern position at the beginning of the Pleistocene (Tugolesov *et al.*, 1985).

Taking into consideration the data mentioned above, the following scheme of seismostratigraphic subdivision of the sedimentary sequence of the Danube deep-sea fan can be assumed. The first (lower) seismic unit SU1 was most likely formed during the Mindel regression and Mindel–Riss transgression, the next unit SU2 during the late Riss regression and Riss–Würm trans-

gression, and the third one (SU3+) during the Würm regression and Holocene transgression.

Accumulation periods of three seismic units were different. For the lower (SU1), middle (SU2), and upper (SU3) units, they were approximately equal to 400, 100, and 30–40 k.y., respectively. The period between the Mindel and late Riss transgressions, both associated, as is mentioned, with the significant sea-level drops, was characterized by the sea-level oscillations of a much lower amplitude (10–20 m). These low-amplitude regressions and transgressions resulted in accumulation of sedimentary bodies that complicate the structure of the deep-sea fan. Precisely these events are responsible for the intricate structure of the transgressive sequence of the lower unit in the fan. In addition to sediments accumulated during the main transgressive phase that followed the Mindel glaciation (the early Paleoeuxinian transgression), the regressive part of the lower seismic unit (SU1) encloses deposits accumulated during several low-amplitude sea-level fluctuations (Paleouzunlarian, late Paleoeuxinian, and Uzunlarian regressions–transgressions) The regressive sequence of the lower unit (R1) is characterized by a large thickness and intricate structure.

The suggested interpretation differs substantially from the previous one published by Shimkus *et al.* (1997). They did not discriminate the high- and low-amplitude regressions and argued that the transgressive accumulative bodies (i.e., those accumulated during the high sea-level stand) are almost absent in the deep-sea fans because, as they believe, periods of normal sedimentation were insignificant and associated with the sharply decreased terrigenous influx.

However, such a situation is not universal. There are examples of thick deep-sea fans accumulated under the high sea-level stand in different seas and oceans: for instance, the deep-sea fan in the central part of the Ionic Sea extending from the Otranto Strait (Moskalenko *et al.*, 1976), or the huge deep-sea fan of the Brahmaputra River that stretches up to the abyssal part of the Indian Ocean (Cochran *et al.*, 1989). The influx of terrigenous material transported by large rivers that have spacious drainage areas and tributaries starting in mountainous regions remains significant as well in the periods of high sea-level stand. In this case, the accumulation areas change position. When the sea level is high, the coarse terrigenous fraction accumulates in the shelf area, however, the finer fraction reaches the continental slope forming the stratiform seismic facies.

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